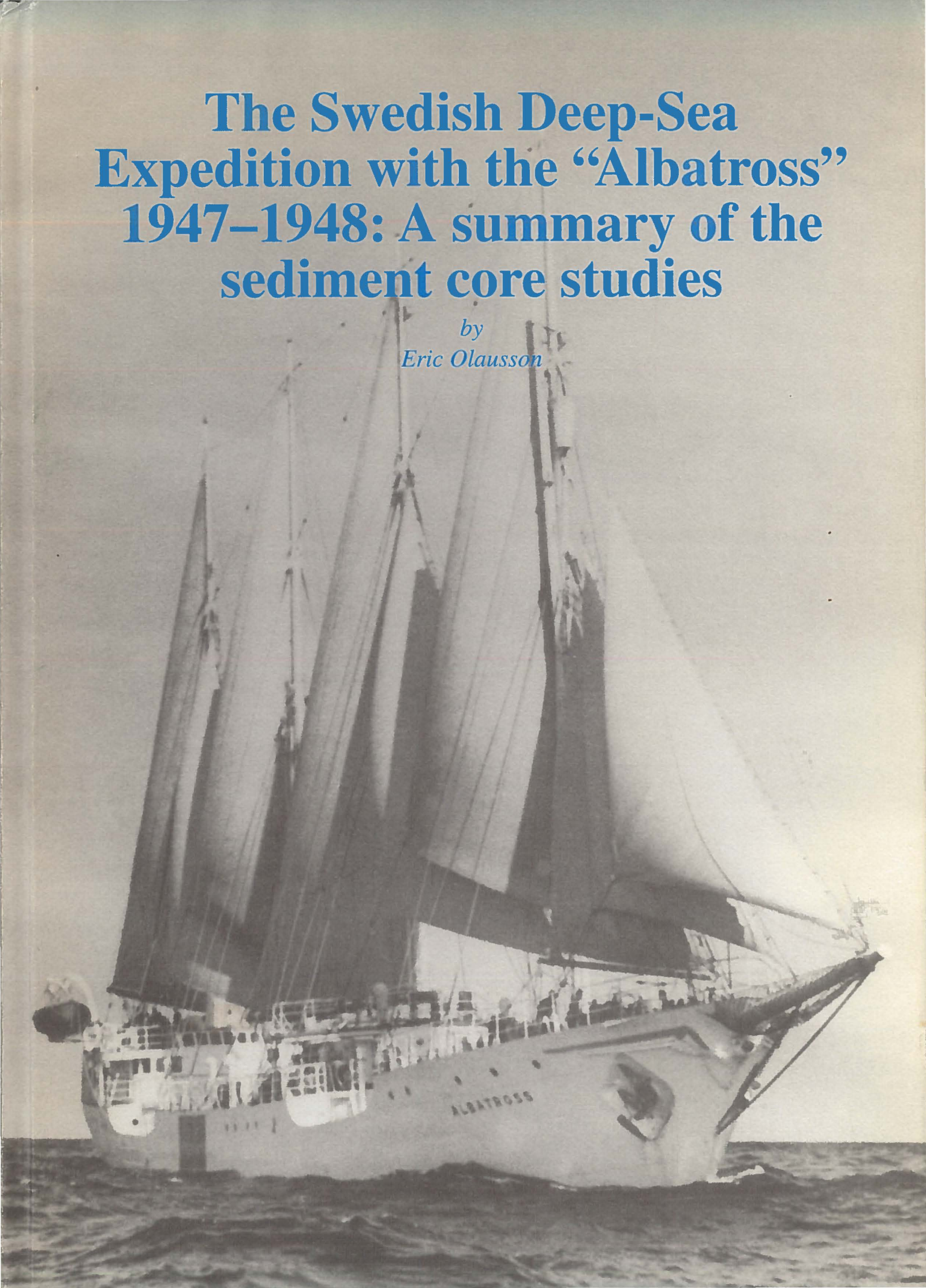
The Swedish Deep-Sea Expedition with the "Albatross" 1947–1948: A summary of the sediment core studies

by

Eric Olausson



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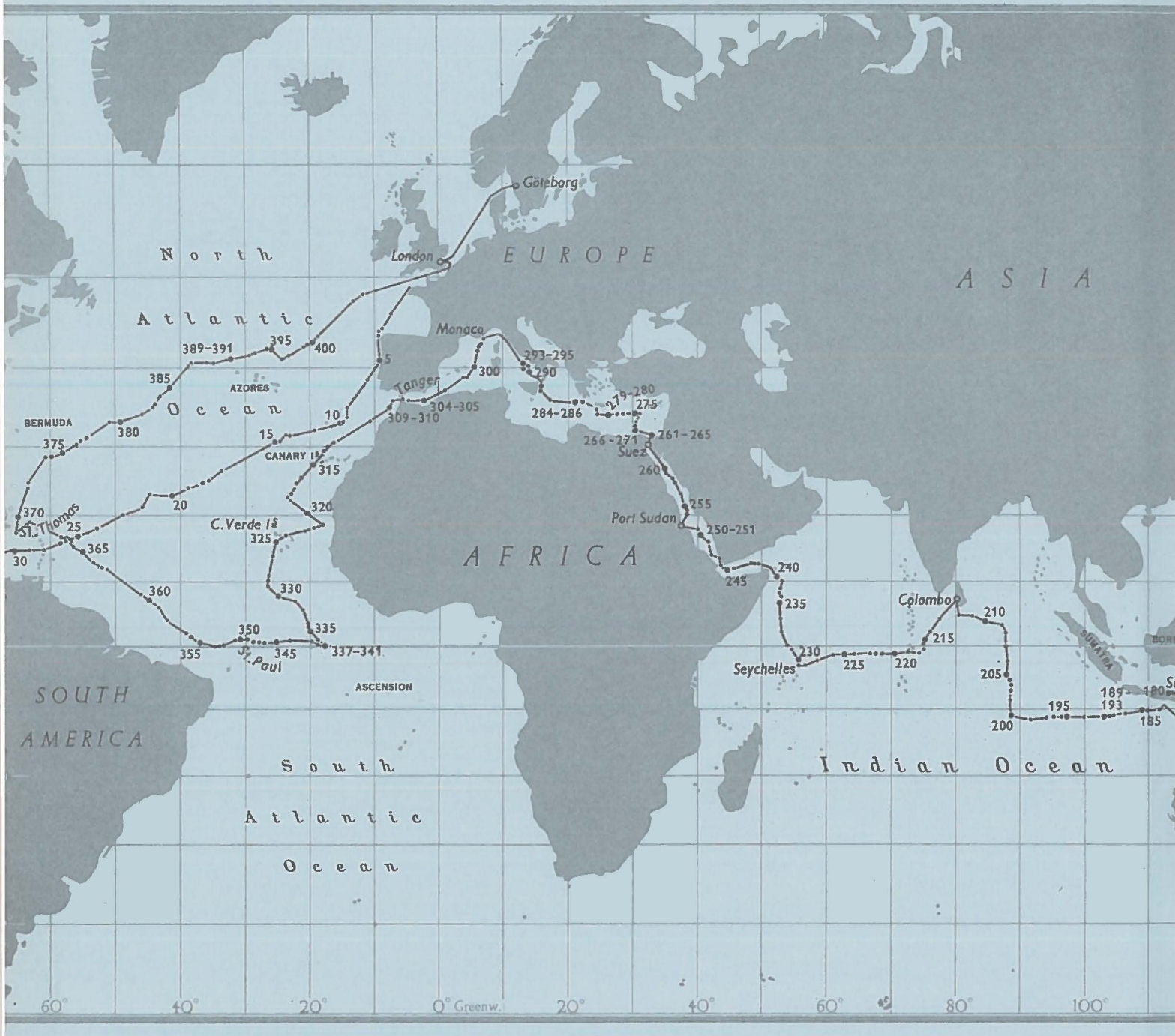

ALBATROSS 1947 - 1948

THE SWEDISH DEEP-SEA EXPEDITION WITH THE "ALBATROSS" 1947–1948: A SUMMARY OF THE SEDIMENT CORE STUDIES

by

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PREFACE

On July 4 1997 it is fifty years since the "Albatross" left Göteborg on its fifteen months long deep sea expedition around the world. During this cruise the scientists on board collected ca 400 long sediment cores with a maximum length of 20 m. The studies of these cores bridged the gap between oceanography and geology. A new science was born: paleoceanography. Instead of studying the ocean´s processes on a time scale of less than a thousand years, the perspective was extended approx. two million years. The complicated interplays between the oceans and their adjacent seas were discovered and found to mirror and influence the climatic changes. By combining these records from the Albatross cores with those obtained from studies of continental deposits, particularly from present and past glaciated areas, a new picture of the Quaternary epoch was established.

It is the aim of this book to sum up the results of the study of the Albatross cores on the occassion of this anniversary celebration and to remind scientists and other people interested in marine science about important Swedish contributions to the study of past oceans.

I dedicate this book to the sponsors of the Albatross Expedition, all from Göteborg, to the scientific staff and the crew of the M/S Albatross, and to my colleagues in the study of the sediment cores collected during the cruise.

Hultvägen 9, S-443 42 Gråbo, Sweden, april 1996
Eric Olausson

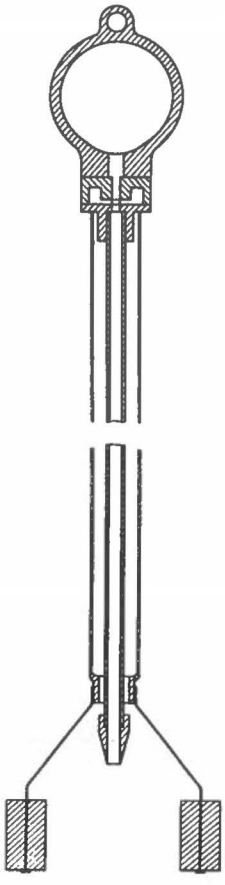
I. The origin of the Swedish Deep-Sea Expedition

Early in this century the radium content of deep-sea deposits was a matter of scientific debate and discussion. Joly (1908) had found that the radium content in abyssal sediment samples was some twenty times higher than in continental rocks. A young student in Uppsala at that time, Hans Pettersson, became especially interested in this element. This interest became deepened during his research work 1911–1912 under Sir William Ramsay at the University College of London. When in London he was invited by Sir John Murray to come over to his "Challenger Office" near Edinburgh in order to use its unique collection of deep-sea deposits for radium studies. However, due to other commitments, Hans Pettersson was not able to accept that invitation, although his interest in problems concerning radioactivity of sediments remained strong.

In 1921 Hans Pettersson and his wife Dagmar, a trained chemist, went to Monaco following an invitation by Prince Albert whose Musée Océanographique had a large collection of deep-sea deposits. However, Pettersson found that the resources available in Monaco were not adequate for making exact measurements. He became invited to the "Institut für Radiumforschung" in Vienna and moved over to that city in 1922. There he was able to carry out more precise measurements than in Monaco. Analysis of the deep-sea deposits from the Monaco collections, later supplemented with samples from the "Challenger" collection, gave results which confirmed the high radium contents earlier reported by Joly. However, none of the explanations so far suggested for the origin of the abyssal radium satisfied Pettersson. To help solve this mystery Pettersson engaged in a decade of sustained efforts in Vienna and in Göteborg. This work involved a detailed study not only of radium but also of its ancestral element uranium in seawater. The final solution was, according to Pettersson (1937), that the high radium content in red clay and in other deposits from great water depths is due to the precipitation of ionium (the immediate parent element of radium) from seawater on to the sea floor where, in the course of several thousand years, its daughter isotope of radium is bred. At that time Pettersson realized the great need for further samples from the deepsea, especially obtained as much longer cores, reaching further back in time than those until then available. The dream of a Swedish deep-sea expedition with sampling of the deep-ocean sediments as its principal objective gradually began to develop.

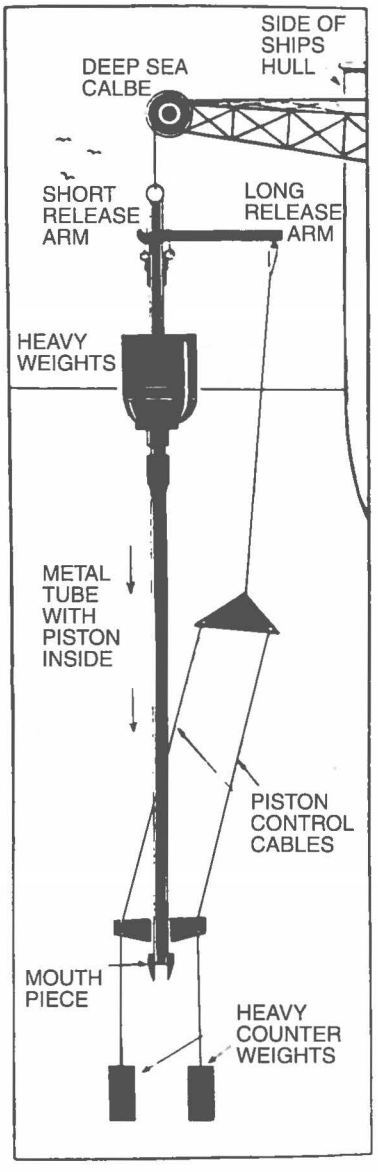
1. The planning and scientific staff (from above and left): Prof. Hans Pettersson (1888–1966), Dr Nils Jerlov (1909–1990), Dr Börje Kullenberg (1906–1991), Dr Fritz Koczy (1914–1967), and Mechanic Axel Jonasson (1903–1991).

In 1930 a young physicist from the University of Lund, Börje Kullenberg, was given a position at the Oceanographic Institute in Göteborg. Kullenberg was well endowed with both technical skill and creative power. At Göteborg he learned about Pettersson´s plan for obtaining long sediment cores by utilizing the hydrostatic pressure at great depths. Together they constructed a sampling instrument called the "vacuum" corer (Pettersson and Kullenberg, 1941). In this instrument, a thick-walled container, fastened to the upper end of a long core barrel made of steel (Fig. 2) is evacuated just before the instrument is lowered to the bottom. A plunger kept in position by strong arrester prevents the seawater from prematurely entering the vacuum container through the core barrel. At contact with the seabottom, the plunger is released and as the heavy coring tube penetrates the sea floor the sediment is forced to rise inside the tube as a results of the high pressure of the surrounding water and the vacuum chamber. The pressure gradient was intended to help the sediment in entering the tube, by reducing the friction against its wall. By means of this instrument a sediment core exceeding 14 metres in length was raised in 1942 from the bottom of the Gullmar Fjord.



2. The vacuum corer.

Kullenberg was not satisfied with the vacuum corer. At the suggestion of Dr Carl Caldenius he started to make a modification of the piston corer used for ground investigations, developed by J. Olsson (1925) at the Geotechnical Department of the Swedish State Railways. In the piston corer (Kullenberg, 1944, 1947, 1955) the piston is being kept stationary at the sediment surface. As the barrel penetrates the sediment, the piston moves upward from the bottom of the barrel, sucking sediment into the barrel. In order to use the free-fall principle a release mechanism was fixed on top of the device, consisting of a hook and a lever. This component was developed by the mechanic engineer Axel Jonasson at the Institute. Jonasson also developed the core catcher at the end of the core barrel. The piston corer (Fig. 3) was the main tool used during the "Albatross" expedition.



3. The piston corer in action.

Pettersson also had another interest. The contact with Dr. W. Weibull of the Bofors Armament Works resulted in the development of a seismic method for measuring the thickness of deep-sea sediments. Dr Weibull was able to work out practical methods for triggering submarine explosions and for recording the echoes from subbottom sediment layers.

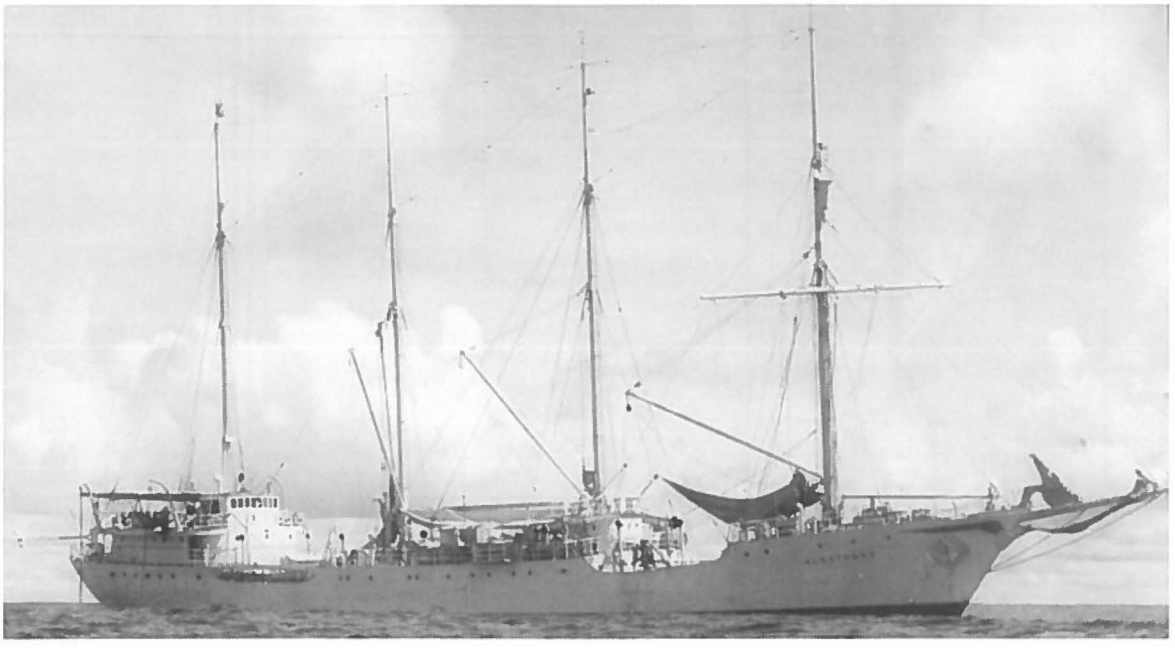
F. Koczy and N. Jerlov developed methods to collect uncontaminated water samples and measure the temperature at accurately determined levels near the deep-sea floor. Jerlov in particular had specialized in developing instruments for the study of optical properties of seawater such as the propagation, absorption, scattering and reflection of light. Pettersson attempted to develop a geothermometer (Fig. 9) in order to study the temperature gradient in the deep-sea deposits. An English firm, Marine Instruments Co., constructed a new echosounder.

Thus, the Oceanographic Institute in Göteborg was now well prepared for a large geological-geochemical expedition. However, a test of the new equipment was found advisable before using it in the open ocean. During a cruise to the western Mediterranean April to June 1946 the RV "Skagerrak" made visits to the Algiers-Provencal and the Tyrrhenian Basins where cores up to 10 m length were obtained. The results of the cruise are summarized in Pettersson et al. (1948).

II. Planning the Around-the-World Cruise

There were two major problems in the planning a world-encircling cruise. Who would provide a suitable ship of the tonnage and the radius of action required? And who would donate the large sum of money for renting and rebuilding of such a ship and for all the expenses for being more than year out at sea? Pettersson (1957, p 22) writes: "The city of Göteborg is second in size and in population to the Swedish capital Stockholm, but it takes the first place among Swedish harbours as a centre of shipping and of shipbuilding. Moreover Göteborg has always been renowned for the generosity of its men of wealth and vision towards arts and sciences". When our foremost Arctic explorer A. E. Nordenskiöld planned his expeditions to Greenland and to Spitzbergen in the 1860´s he turned to Baron Dickson of Göteborg for financial backing, and promptly got it. Thirty years later S. A. André, when planning his ill-fated attempt to reach the North Pole by balloon, also found his main supporters in Göteborg. The generosity towards oceanography shown by leaders of industry and commerce in Göteborg such as G. Ekman and K. Mark also must be mentioned. K.A. Wallenberg supported the Gullmar Fjord research stations on Bornö and in Lysekil. Wallenberg later on also decided to give the large sum required for building and equipping the Oceanographic Institute.

The president of the Broström Combine at that time, Major H. Jacobsson, offered the new training ship of the Broström Combine, the motor-schooner "Albatross" of 1,450 tons (Fig. 4). He gave permission to have it outfitted with laboratories, workshops, cabins and other accomodation for scientific and technical staff. In addition he and his wife made a large donation of 500,000 Swedish crowns (SEK) to help finance the expedition. Another promoter of the plan was the County Governor Malte Jacobsson of Göteborg. Largely owing to his recommendation the greatest textile magnate of Göteborg, Gustaf Werner, had earlier promised 1 million SEK for sponsoring the cruise. Finally, the Consul General A. Axelsson Johnsson, owner of the Lindholmen shipyard in Göteborg, granted Pettersson a considerable reduction in the estimated costs for the rebuildning of the ship. – When correcting for inflation, the value of the donated 1.5 million SEK would be in about 20 million SEK in 1995 (vs 3 million US dollars).



4. The "Albatross"

III. The Members of the Scientific and Technical Staff

Professor Hans Pettersson was the principal scientific leader of the expedition.

Dr Börje Kullenberg was the head of the coring operations, while

Dr Nils Jerlov was the leader of the hydrographic and the optical studies of the sea water.

Dr Orvar Nubelin performed biological studies in June 1948.

Professor Waloddi Weibull was supervisor of the seismic work in July–August 1947.

Dr Fritz Koczy dealt with geochemistry, but he also was in charge of the echo-soundings.

Dr Gustaf Arrhenius was the head of the core laboratory.

The chemist Leif Bruneau carried out the water analyses until the end of May 1948, when his job was taken over by another chemist, Nils Karlsson.

Engineer Viggo Wentzel performed the seismic works and studies of the ionosphere.

Axel Jonasson was the principal engineer, assisting Kullenberg in the coring operations. Being in charge of the deep-sea winch operation, he took part in trawling and he was further responsible for upkeep and repair of the scientific instruments.

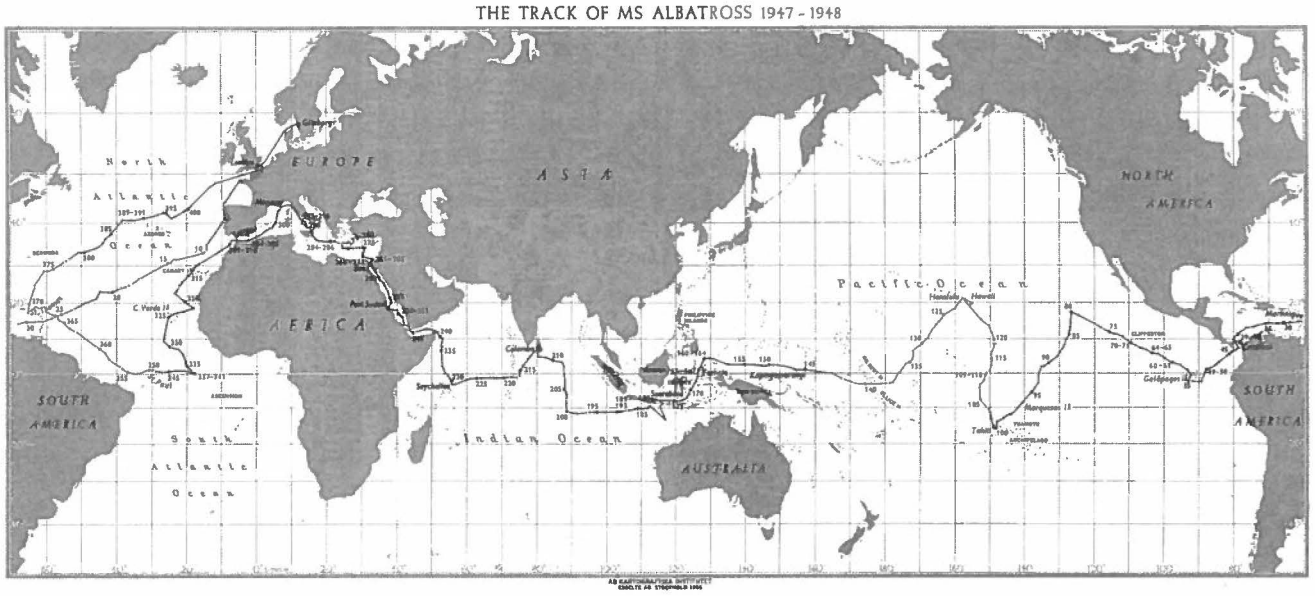
Mr Rutger Irgens (Hans Pettersson´s son) and Harry Quist assisted in sediment curation and in mechanical work in general.

Professor Fred Phleger, a micropaleontologist, took part while the expedition was in the Carribbean.

Dr John Eriksson was the surgeon and naturalist of the expedition.

The 19 men large crew of the "Albatross" were managed by Captain Nils Krafft and the Chief Engineer Hugo Enwall. In addition, 12 trainees took part in the expedition.

IV. The Expedition July 4, 1947–October 3, 1948



5. The track of the M/S Albatross 1947–1948

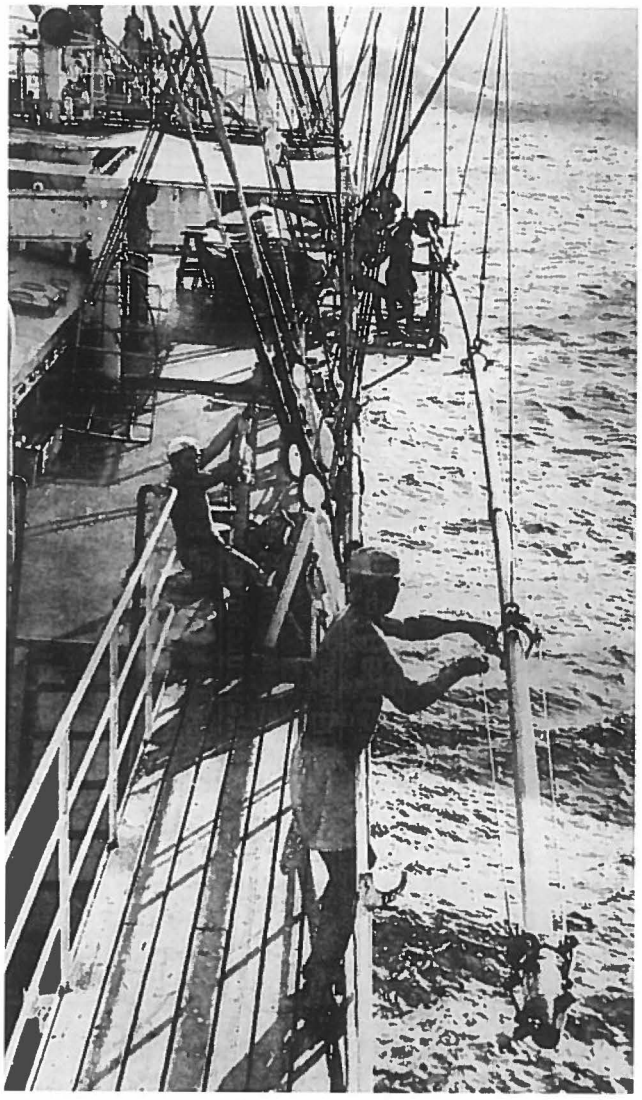
1. The Atlantic I

On July 4th 1947 the "Albatross" left the harbour of Göteborg (Fig. 5). Pettersson describes in his "The Voyage" (1957) their first attempts to take cores and the repeated problems with the winch. He had expected to find a smooth topography in the great depths but found that plane and smooth areas extending for hundreds of kilometres were rather an exception than a rule. The sea floor topography of the oceans was originally based on wire soundings which gave a simple impression of a smooth ocean floor with a broad midoceanic ridge. However, the introduction of acoustic soundings, used e.g. by the German "Meteor" in the twenties had revealed a much more irregular relief in large parts of the oceans.

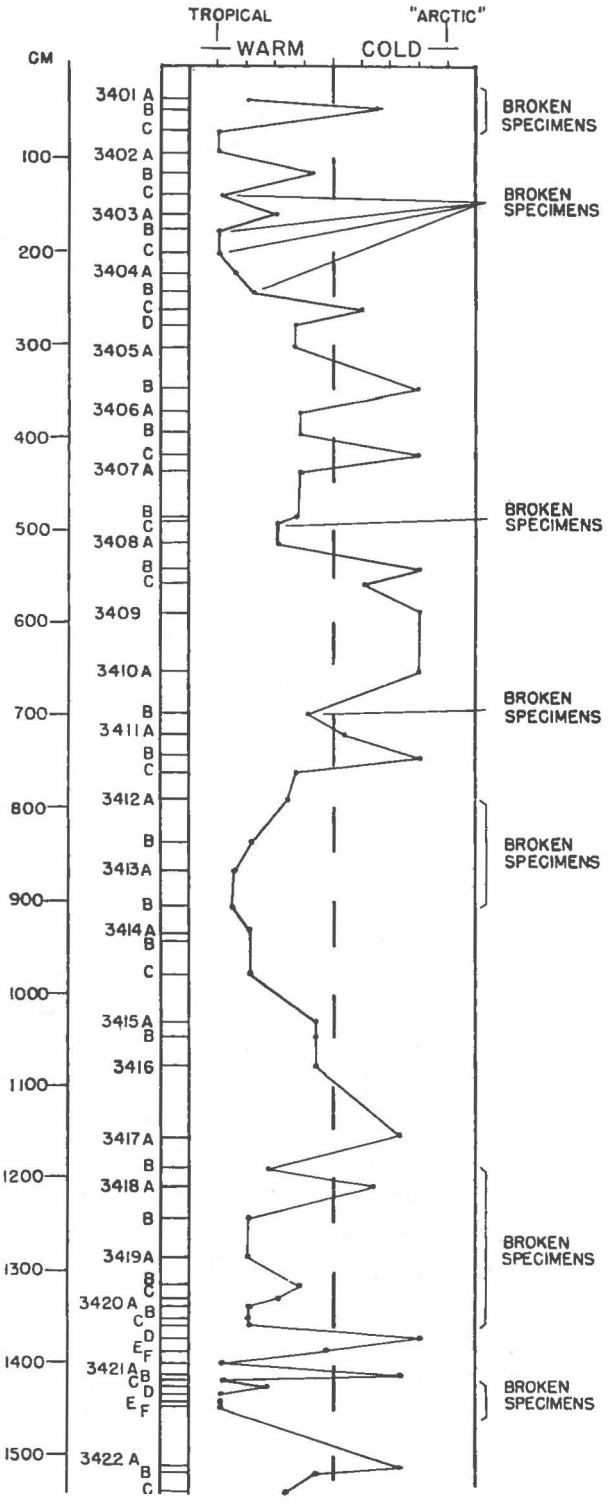
Fifty years ago, the formation of the ocean basins was a subject for speculations with practically no direct evidence available. The problem at that time was whether or not the ocean basins always have shown the same dimensions and shapes as they do at present. Scientists such as Umbgrove and Wegener had given different views, as had Du Toit. Weibull´s and Wentzel´s attempt to measure the thickness of the sediment cover in the Canary Basin were difficult to interpret, but the subbottom echoes at 3,475 m suggested that the time span required for the formation of this supposed sediment top layer should be of the order of at least 500 million years or possibly twice as much as that. Today we know that the oldest sediments at these sites are only Upper Cretaceous in age (60–90 million years).

A lower subbottom reflector was later found at a sediment depth of 300 m in the western Atlantic. With the same rate of accumulation as assumed for the Canary Basin the age of the western Atlantic reflector may be Early Tertiary i.e. a good estimate also in the light of the present knowledge.

In the Caribbean Basin a 15.4 m long core was raised, studied by Fred Phleger (1948). After that, the winch was out of order until it was repaired in Panama. Phleger´s pioneering foraminiferal stratigraphy of this Caribbean core indicated a number of warm-cold cycles in the region. The 16 planktonic foraminiferal species found in the core were interpreted in terms of relative water temperatures. High



6. The piston corer is taken on board. The man with the corer is Axel Jonasson.



7. Foraminiferal diagram of Core 34 from the Caribbean.

frequency of *Globigerina bulloides, G. pachyderma, Globigerinoides ruber* and absence of *Globorotalidae* were taken to indicate cold climate or ice ages while high frequency of *Globorotalia menardii, G. tumida, Pulleniatina obliquiloculata, Globigerinoides sacculifera,* and *Globigerina eggeri* were seen as markers for warmer conditions (interglacials). Using this method (afterwards employed by many other scientists), Phleger distinguished four glacial units (Nebraskan, Kansan, Illinoian and Wisconsin) and three interglacials (Aftonian, Yarmouth, Sangamon) in this core (as did subsequently by Pettersson, 1950). Thus, we now the first well-established deep-sea stratigraphy extending back half a million years (Fig. 7). Earlier attempts had only reached the upper part of the last ice age, i.e. Wisconsin in American terminology (Weichselian in Northern European and Würm in the Alpine stratigraphy).

The cores collected during the trip from Lisboa to St. Thomas, nr 21–36 were sent back to Sweden from Panama. A few of these were studied, particularly by W. Schott (1952, 1966). These cores will be discussed later on together with the other cores.

2. *The Pacific Ocean*

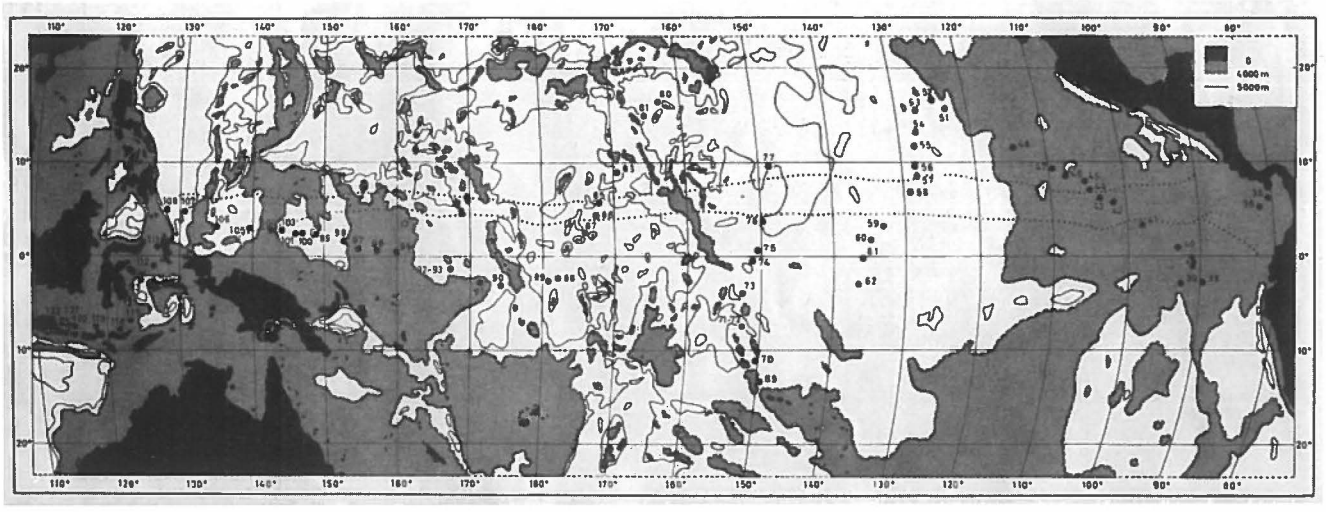
The original plan was to make four cross-transects through the Equatorial current system in the Pacific:

Section 1: the cores 35–36 from the Panama Basin and the cores 38–52 from the transect from about S 03°,W 90° to N 16°, W 125° (Fig. 8). These cores were studied by Arrhenius (1952).

Section 2: from N 16°, W 125° down to Tahiti: the cores 53–69.

Section 3: Tahiti to Hawaii: the cores 70–77.

Section 4: Hawaii to Gilbert Island: the cores 80–88.



8. Location of the cores from the Pacific. The dotted lines indicate the limits of the Equatorial Counter Current in February–March.

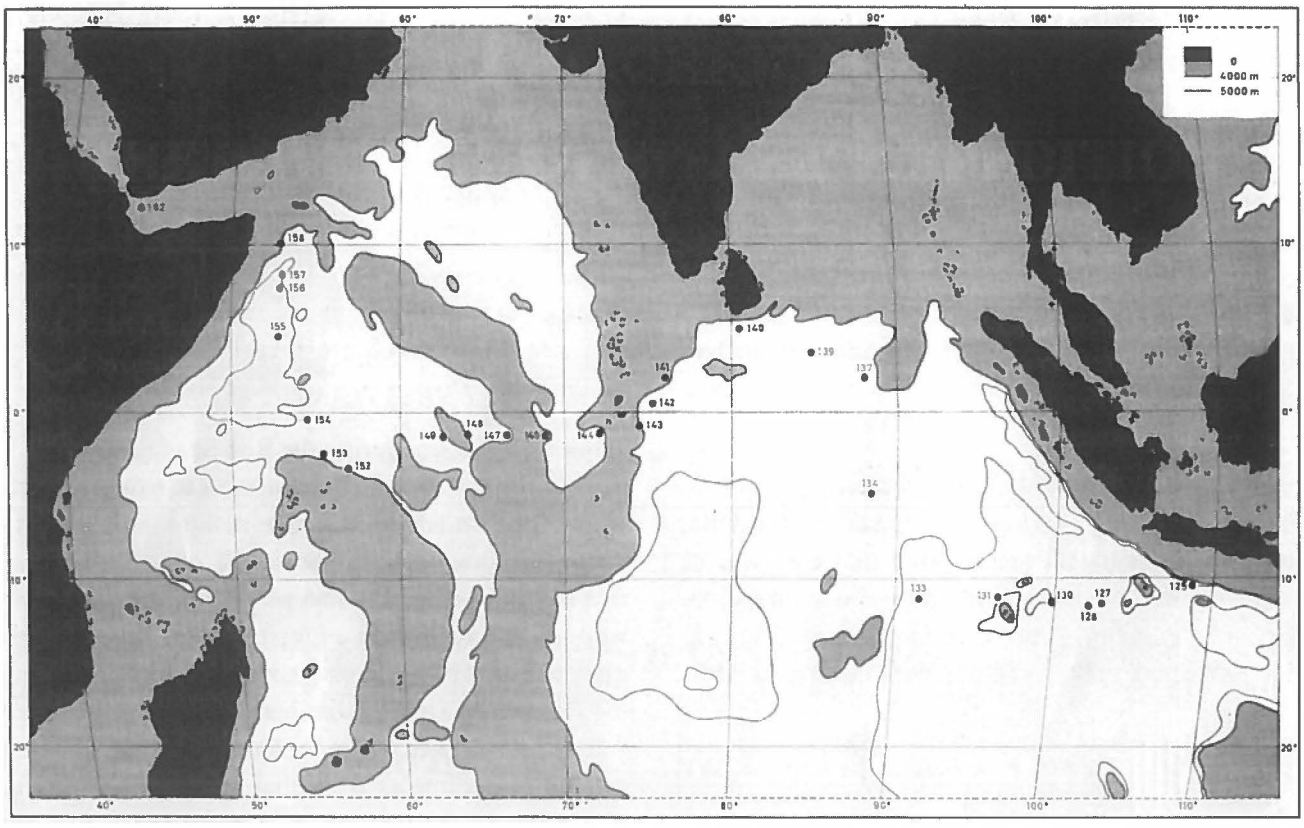
Based on previous experience with damage to coring equipment, the length of the corer was limited to 10.5 m or only 5.5 m where the bottom conditions appeared unfavourable. However, attempts to use a 15 m long corer also were made and were successful in the central and western Pacific. Experiments with a "geothermometer" (see Fig. 9) at the equator, W 150°, gave an unexpectedly high geothermal gradient, 1°C per 21.5 m. According to present knowledge this is not high; a gradient twice as large is not uncommon. Thus, even with his primitive instrument, Pettersson was on the way to discover the large heat fluxes that locally characterize the oceanic lithosphere.

Of all cores of the expedition the one from the greatest water depth was collected in the Mindanao Trench, at a depth of 7785 m. The mud in this core are described by K. Fredriksson et al. (1958). A series of graded sand layers were interpreted as turbidites in the sence of Kuenen.

9. Axel Jonasson and the geothermometer he made.

3. *The Indian Ocean*

The easternmost part of the Indian Ocean is in bad repute amongst sailors since it is often visited by tropical storms. When she had left the Isle of Beauty, "Albatross" ran into a severe gale, the worst of the entire cruise. During that storm the ship rolled to nearly 36° to either side; the height of the waves gradually increased to about 8 m. The captain had to change the course by 180° to get away. Five days later, the wind had abated and a day after coring operations started again.



10. Location of cores from the Indian Ocean.

Coring operations in the Indian Ocean mainly involved two sections across the equatorial current systems, one along E 88°, the other along ca. E 53° (Fig. 10). Cores were obtained both from oligotrophic areas and from more productive regions e.g. below the South Equatorial Current and the (seasonal) South-west Monsoon Current.

4. *Red Sea*

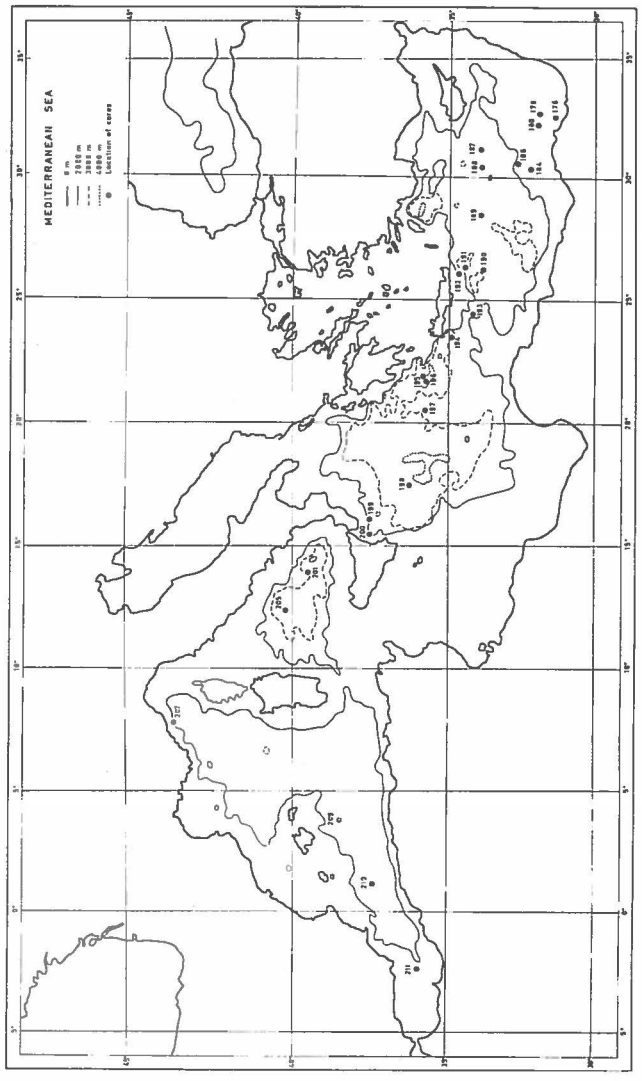
Not all members of the expedition spent all their time on board the "Albatross". On the Seychelles Gustaf Arrhenius temporarily left the expedition to visit Sweden; he rejoined it later in Suez. Pettersson also made a trip to Sweden. He left in Port Sudan and returned in Monaco. Pettersson reports in "The voyage" that the cruise through the Red Sea was "uneventful". However, the "Albatross" discovered unusually warm and salty bottom water at Station 254: at a depth of 1930 m a water temperature of 24.53°C and a salinity of nearly 44 per mil were recorded (Fig. 11). If the hydrocast had been taken 10 nautical miles more to NW, they would have sampled the hot brine, discovered in 1967 by the Atlantic II Expedition (Degens and Ross 1969). It has been said to me that the scientists on board "Albatross" did not trust the results

11. Location of the cores from the Red Sea. Brines have been found in median valley, both to the north and south of the station where core 168 was obtained.

from the bottom water samples at the station 254. A confirmation of the results would have been obtained if they had made a further survey of the axial valley of the Red Sea. The sediment cores collected in the Red Sea were rather unusual, according to Pettersson, containing large aragonitic concretions and giving off a distinct smell of H_2S. That would have aroused the interest of a marine geologist. However, the fact that both geoscientists (Arrhenius and Pettersson) were absent during the cruise in the Red Sea may answer the question why no further studies were made there.

5. *The Mediterranean*

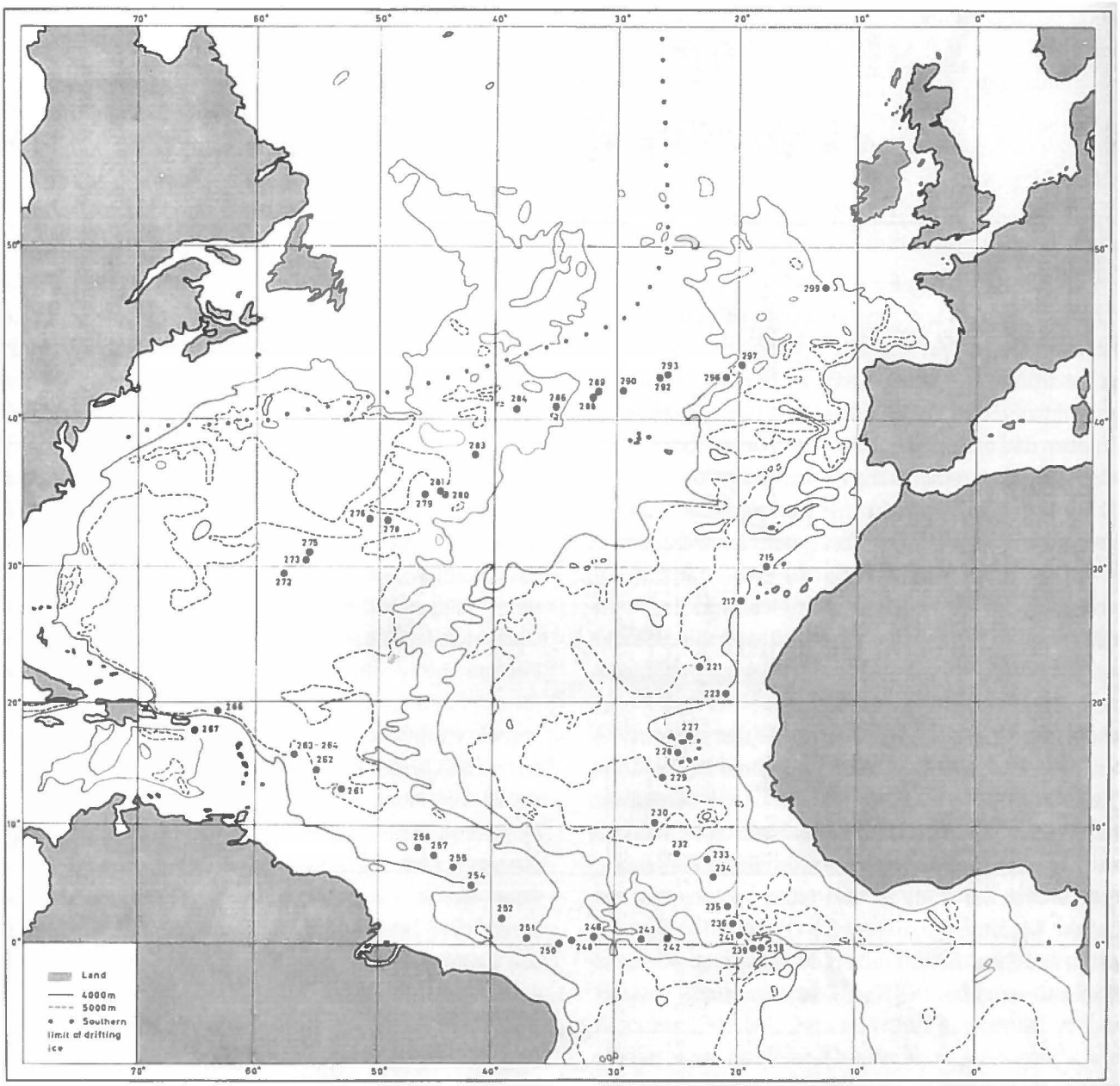
The coring started in the region of the outer Nile delta, but the stratigraphical record from that area was of limited extent due to the high rate of accumulation. Farther to the north, from Cyprus to Sicily (Fig. 12) the "Albatross" collected cores with many dark sediment layers (sapropels), suggesting drastic changes in bottom water conditions. Also many layers of volcanic ash were noticed. The cores from the western Mediterranean were more sandy and here only very few sapropel layers were found. Volcanic matter was abundant in the Tyrrhenian Sea. Off central Italy, not far from the volcano Vesuvius a 18 m long core was taken. This core later on was studied by Erik Norin (1956).



12. Location of the cores from the Mediterranean.

6. *The Atlantic Ocean II*

While in Göteborg during the Red Sea leg of the "Albatross" Pettersson had found means to extend the expedition by an additional three months, in the North Atlantic. Cores were collected off northwestern Africa to study the past eolian activity of the Sahara, as well as the formation of certain minerals such as glauconite. The "Albatross" spent several days around the Romanche Deep. From here a core was obtained suitable for studies of the mineralogy and the radium content of sediments from that deep. An attempt to visit St Paul´s Rock failed. Four important cores were collected during the cruise while on the way to the equator: nrs 234, 241, 242 and 246. This will be discussed later on. During the last transect, from Bermuda to London (Fig. 13), several long and stratigraphically important cores were raised as well. Particularly Core 280, reaching back more than half a million years in time, became famous of the $\partial^{18}0$ studies by Emiliani (1958). On September 23 the "Albatross" arrived at the port of London, and 6 days later she left London, and arrived in Göteborg October 3, 1948 with its unique harvest from the ocean depths. In its cooling storage there were 265 cores, a cargo such as had never been delivered before; 34 cores were already in Göteborg, sent home from Panama. Also water samples and all kinds of analytical data obtained during the cruise was brought back for further treatment and interpretation. Pettersson and the entire scientific party were aware that the material collected during the 15-month cruise was so vast that international co-operation was necessary to get it completely worked up and analyzed.



13. Location of the cores from the North Atlantic.

V. PLANNING THE LABORATORY ANALYSES OF THE CORES

The large collection of deep sea cores, a total of 299, with a combined lenght of 1600 m, was the most important matter brought back from the cruise around the world. So far as I know – I only became involved in the study of the Albatross cores in October 1956 – all cores were opened, briefly described and sampled. Every section of each core, 60 cm in length, was cursorily analyzed and samples were taken every 10 cm. These samples were sent to Dr Olof Arrhenius (the father of Gustaf Arrhenius) who analyzed the ca.16 000 samples as to their content of $CaCO_3$, C_{org}, P, and N. In this way the sediment cores and their composition were generally characterized. Based on this information, regional and special studies could begin. I have been told that responsibilities were to be assigned on a regional basis. Thus, Gustaf Arrhenius was to take care of the cores from the Eastern Pacific, Hans Pettersson to direct the study of the Western Pacific cores, Börje Kullenberg to take over the study of the Indian Ocean, Red Sea and the Mediterranean, and Dr Otto Mellis, a geologist at the Stockholm University, was to be responsible for the study of the North Atlantic cores.

Hans Pettersson had very good contacts in the scientific world, which proved extremely valuable in getting co-operation in the study of the treasure of cores now housed at the Oceanographic Institutet in Göteborg. Some 30 scientists from all parts of the world took part in the analysis of the cores, and many more requested (and still request) samples for special studies. Two of these scientists from abroad, who did fundamental work on the Albatross cores, were the outstanding scientists Fred Phleger and Cesare Emiliani. Phleger had taken part in the expedition for the Caribbean leg. Together with Frances Parker and Jean Peirson he produced an excellent study of the foraminifera in the North Atlantic cores. Dr Parker and Dr Ruth Todd made similar studies of the Mediterranean cores. During a visit to the Enrico Fermi Institute for Nuclear Studies, University of Chicago, Pettersson came in contact with Emiliani. It was this contact which resulted in the first papers about isotopic ocean paleotemperatures. These studies included five Eastern Pacific, three North Atlantic and one Eastern Mediterranean core.

VI. A NEW SCIENCE IS BORN: PALEOCEANOGRAPHY

Up to the time of the Swedish deep-sea expedition, geology and ocenanography were quite separate scientific disciplines. Contacts was minimal – with a few isolated exceptions (e.g. W. Schott of the Meteor Expedition). Geologists dealt predominantly with the geology on the continents and sedimentation on shelves, while oceanographers studied processes in the present ocean. When the "Albatross" came back with some 1,600 m of sediment cores, she carried with her the arcives of ocean history and of ocean-climate interaction. New methods could now be developed for reconstruction of past circulation, ocean chemistry and productivity through time. From the study of the "Albatross" cores, during the 50´s and 60´s, a new science was born: paleoceanography, based on the combination of methods and theories from geology and oceanography. "The marine geologists working on these cores, thus, established the new field of Earth Science that we now call ´paleoceanography´. Gustaf Arrhenius produced evidence for significant changes in productivity in the eastern tropical Pacific, and argued for corresponding changes having occurred in the strength of trade winds; Cesare Emiliani established the cyclicity of the ice-age oxygen isotope record, and argued for the importance of Milankovitch forcing in climatic change; Eric Olausson proposed that carbonate preservation patterns contained clues to changes in deep-sea circulation; Frances Parker and Fred B. Phleger produced evidence for large – scale shifts in climatic zones in the last hundred thousands years. The ideas of these pioneers have guided much research since, especially in Quaternary studies" (Berger, 1994). Their methods and the new findings added during the 60´s, particularly paleomagnetism and plate tectonics, are still the main tools in marine geological studies. Let us start the discussion about the methods used on the cores of the Swedish Deep-Sea Expedition with the new isotope methods.

1. THE ISOTOPIC PALEOTEMPERATURE ANALYSIS

The application of oxygen isotope ratios in fossil calcareous solids to reconstruct ancient temperatures was first proposed by Urey (1947) at the University of Chicago. In order to attain sufficient precision in the measurements he and his collaborators constructed a new mass spectrometer (Nier 1947). McCrea, also at the University of Chicago, developed a method of extracting CO_2 from $CaCO_3$ (McCrea 1950) and Epstein (1951) solved preparation problems for shells and fossils. Thus, a routine analytical method for isotopic paleotemperature analysis was available a few years after the Swedish Deep-Sea Expedition. During a visit to Chicago in the early 50´s, Hans Pettersson met Cesare Emiliani who worked in Urey´s lab, analysing samples from the Lomita Marl, California. Emiliani asked Pettersson if it was possible to get material from his deep-sea cores. "That is how the whole thing started" (Emiliani, priv.comm.). Emiliani got samples from the Pacific cores 53, 57, 58, 60 and 61, from the Eastern Mediterranean core 189, and from the Atlantic cores 234, 241, 246, and 280. The results of these studies were published by Emiliani a few years later. Emiliani´s paper "Pleistocene temperatures (1955a) is a classic paper. The other two papers dealing with the Albatross cores are Emiliani (1955 b) and (1958).

The isotopic temperature analysis is based on the determination of the ratio of the oxygen isotopes $^{18}0$ and $^{16}0$ in the calcium carbonate of fossils, in the case of deep-sea sediments, foraminifera. This ratio is a function both of the temperature and of the isotopic composition of the water in which the foraminifera lived. Therefore, it is desirable to know both the depth habitats of the planktic foraminifera as well as the isotopic composition of the ambient sea water, as is clear from the empirical equation

$$T = 16.5 - 4.3(\partial_{sample} - \partial_{sea\ water}) + 0.14\,(\partial_{sample} - \partial_{sea\ water})^2$$

where T is temperature, ∂_{sample} and $\partial_{sea\ water}$ are, respectively, the $^{18}0/^{16}0$ indices of the analyzed carbonates

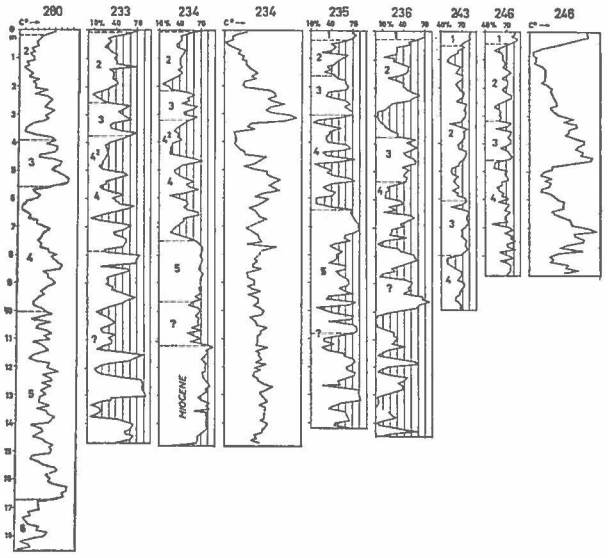
(foraminifera) and of the water where they grew (Epstein et al., 1953). However, only the isotopic ratio in the sample (carbonate) can be measured by the mass spectrometer analyses. Depth habitat and water composition must be estimated or will remain uncertain. The ∂ value is given as a proportional deviation of isotopic ratios from a standard:

$$\partial(m) = (R_{sample}-R_{standard})/R_{standard},$$

where R is the ratio of the heavier to the lighter isotope and m is the index of the sample. Values for ∂ are given in permil, that is 1000 times the actual value of the deviation. The standard chosen is an international one to which all analyses in all laboratories refer to.

When Emiliani first analyzed planktic foraminifera, several hundred tests were required to make up the 5 mg of calcium carbonate necessary for each determination. Fortunately modern mass spectrometers can analyze much smaller samples, down to a single test. The planktic foraminifera have different depth habitats. Emiliani (1954) found that some species (e.g. *Globigerinoides ruber* and *G. sacculifera*) live closest to the sea surface, while others (e.g. *Globigerina bulloides, Globigerina inflata, Globorotalia tumida, Pulleniatina obliquiloculata*) occupy deeper habitats. By selecting tests of appropiate species, information about the temperature and/or the isotopic ratio of a certain water layer could be reconstructed. The species living close to the sea surface, e.g. *Globigerinoides sacculifera,* are usually preferred since surface temperatures are the most important item in climate reconstruction. Deeper living foraminifera are analyzed when the water structure is to be reconstructed or in cases in which "surface species", usually thin-walled, are not available either because of dissolution or because of unfavourable conditions for growth.

Emiliani (1955 a) realized that the composition of seawater must change through time as a function of ice mass present on Earth. Thus, before temperature can be read from the $\partial^{18}0$ values a correction has to be made for the ice effect. He thought that the northern ice sheets (now vanished) had a $\partial^{18}0$ of about – 15 per mil. This works out, by mass balance, to a correction near + 0.5 per mil for the ocean water during glacial ages. The isotopic range between glacial and interglacial, diminshed by 0.5 per mil, would then yield the temperature range. In this way Emiliani presented a "temperature" amplitude for the Quaternary period of about 4°C for the tropical Pacific, up to 10°C of the North Atlantic (Fig. 14). For the Eastern Mediterranean (Core 189) very large amplitude was given, up to 25°C, now known to be erronous due to neglect of meltwater effect during stagnation (see further Chapter XII).



14. Isotopic paleotemperature and calcium carbonate curves of some Albatross cores from the Atlantic Ocean (280, 234, and 246). Stages and stage boundaries are marked. Stage 1= Holocene, 2 = Weichselian maximum, 3 = Early Weichselian and Eemian (isotopic stages 4–5), 4 = Saalian (isotopic stages 6–10), 5 = Holsteinian (isotopic stages 11–15) , and 6 = (Late) Elsterian (isotopic stage ≥16). Redrawn from Emiliani (1955) and Olausson (1965).

During my studies of the stratigraphy of the "Albatross" cores I became aware of problems with the isotopic paleotemperature analysis. This critical attitude became strengthened when I read a paper by Broecker et al. (1960) with the title "Evidence for an abrupt change in climate close to 11,000 years ago". That was the signal to me that something was wrong in the interpretation of isotopic data. The meltwater effect had been ignored! In Progress in Oceanography, vol. 3, dedicated to Professor Hans Pettersson by his former students and associates on the occasion of his seventy-fifth birthday, I made some remarks on isotopic paleotemperature analysis pertinent to this question. I proposed that the isotopic compositions of the present and ancient ice sheets were about twice as large as that presumed by Emiliani (that is – 30 o/oo instead of – 15 o/oo). Thus, the correction for the waxing and vaning ice sheet had to be considerably greater. I proposed ca. 1.2 per mil (Olausson 1965). Thus, two thirds of the isotopic "paleotemperature" variation in fact was attributable to changes in the

composition of seawater and, only about one third to the temperature variation of the water. I realized then that the temperature amplitude at lower latitudes during the late Pleistocene had to be much less than stated by Emiliani. The amplitude generally increases with latitudes. The ∂18 changes found in cores are ultimately climatically determined. To what degree given peaks or valleys in a given isotopic curve are due to changes in the ∂18 of sea water or due to changes in temperature is mostly unanswered. Closer to the formerly glaciated areas, the meltwater effect, (denied at first by Emiliani and others), is now accepted as a reality. The meltwater effect (meltwater spikes) will be discussed later on in connection with the Mediterranean (Chapter XII) and the cores from northernmost Atlantic in Chapter VI:3 C.

For the last several decades all published isotope data are given as ∂18 relative to the international standard PDB (Pee Dee Belemnite). No longer is "isotopic temperature" offered as an index, as in Emiliani (1955a) However, the ∂18 analysis of foraminifera introduced by Emiliani has been and still is a very important tool in marine geological studies.

I have performed several complete isotopic analyses of the Albatross cores using my own facilities at my Marine Geological Laboratory, from 1980 the Department of Marine Geology, University of Göteborg.. Some of the diagrams are published by me (see e.g. Fig. 41, 42, 44, 45, 57, 64, and Olausson, 1983). Other analyses are in diagrams placed at students' disposal and used in their papers (see e.g. Bornmalm, 1988 , and Naidu et al., 1993).

When I began to interpret the Albatross cores in climatic terms there existed two ways. Arrhenius (1952; see Fig. 18) only used numbers (1–18) for his Pleistocene sequences, i.e. the last million years. Phleger (1948) and Emiliani (1955 a) had correlated their climatic curves with continental stratigraphies. Besides, Emiliani also used core stages (1–14) for the last 300,000 years. Later his core stages became isotopic stages. I found Emiliani´s time-stratigraphic attempt not realistic, particularly his correlation of stage 6 (a single valley) with the Riss ice age and stage 10 (also a single valley) with the Mindel. These two large ice ages were certainly divided into two–three phases as did the Würm ice age and, further, the durations given by Emiliani were too short for so large cool periods. I understood that Arrhenius interpretation was a more realistic one but instead of only numbering I preferred to use the European stratigraphy *sensu* Woldstedt. My climatic curve (Olausson 1961 b), based mainly on carbonate curves, was apparently accepted e.g. by textbook authors as Butzer (1964), Woldstedt (1969) and Flint (1971).

As appears at Fig. 14 and 33 (Olausson 1965, 1971 a) I tried for a short period to use the conventional stratigraphic numbering system: starting from the Holocene every interglacial got an odd number and every complete ice age an even number. However, it brought me into difficulties: what is interstadials and interglacials, what is only an ice advance phase and what is a complete glacial age between two interglacials in Pre-Saalian times? These problems are still unsolved in the isotopic stratigraphy. Since I early understood that the climate in the equatorial part of the Indo-Pacific Oceans had been rather constant during the Pleistocene the cause of the changing content of carbonate, silica etc. as well as the variation of the $\partial^{18}0$ in foraminifera observed in their deep-sea sediments are to the main the results of changes elsewhere. It must be of great interest to use a stratigraphy which informs about the hydrographic and climatic development in the source area. The Indo-Pacific Oceans may be considered as a sink while the Nordic Seas-Arctic Ocean is the source in close interplay with the nearby parts of the continents. As will be discussed in Chapter XIII the size of the glacial ice sheets over the North America has been rather constant since the Nebraskan ice age. This may be because the source of moisture (from the Pacific) has not changed. Therefore the North American terminology is not of great interest to marine geologists. On the other hand it is necessary for a globally interested marine geologist to be well informed about the background of the North European Pleistocene stratigraphy, the development of the Nordic-Arctic Seas, the former extent of ice cover over the Eurasia, deepening of the Bering Strait, and the interplay between land and sea there. These items will be illustrated in Chapters VI:3, X–XIII.

I know very well that marine geologists routinely use the new isotopic stratigraphy. This has largely replaced the European and American continental stratigraphy. However, the continental nomenclature should not be abandoned because of the work that has gone into sorting out ice age history on land, and the necessity of trying marine and land records together. For these reason I have tried correlation between Northern European continental and the marine isotopic and carbonate stratigraphies in the following

way (the Alpine and North American equivalents within parentheses; ka = thousand years, Ma = million years):

Holocene = isotopic stage 1
Weichselian (Würm, Wisconsin) ice age = 2–5d
Weichselian II = 2
Weichselian I/Weichselian II (Oerel- Denekamp) interstadials = 3
Weichselian I = 4
Odderade interstadial = 5 a
Brörup/Odderade stadial = 5 b
Brörup interstadial = 5 c
Eemian/Brörup stadial = 5 d
Eemian (Riss-Würm, Sangamon) interglacial = 5e
Saalian (Riss, Illinoian) ice age = 6–10
Warthe stadial = 6
Rügen interstadial = 7
Drenthe stadial = 8
Treene interstadial = 9
Saale 1 stadial = 10
Holsteinian (Mindel-Riss, Yarmouth, and the Great) interglacial = 11–15
Holsteinian s.s. = 11
Kaluga – Fuhne = 12
Oka – Anglian = 14
Elsterian (Mindel, Kansan) ice age = 16–20 (22)
Brunhes/Matuyama boundary = base 19;792 ka ± 10 k.y.
Cromerian (Günz-Mindel, Aftonian) interglacial = 21 (23) –
Menapian (Günz, Nebraskan) ice age = 26–28
Jaramillo event = mid 27–mid 31; 0.99–1.07 Ma.
Waalian interglacial
Eburonian ice age (≈ 1.4 –1.6 Ma)
Tiglian interglacial (1.6–2.2 Ma)
Pretiglian ice age
Olduvai event = 63–71; 1.77–1.95 Ma
Reuverian (latest age in the Pliocene or > 2.8 Ma).
Matuyama/Gauss bounday. 2.60 Ma

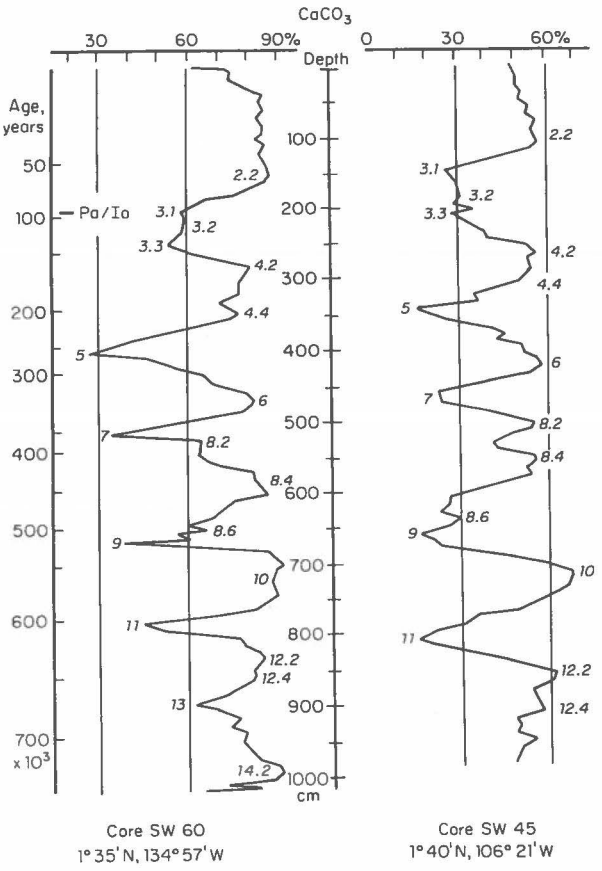
The ages given above are from Berger et al. (1994). As to the correlation between marine and continental stratigraphies see also Fig. 72, and Zubakov and Borzenova (1990, p. 375).

2. *The calcium carbonate variations*

Early deep-sea expeditions such as the Challenger Expedition and later the Carnegie 7th Cruise had revealed the important relationships between the carbonate distribution pattern on the sea floor and that of depth and circulatory pattern (Murray and Renard 1891, Revelle 1944). In particular, it was noticed that the carbonate content of the deep sea deposits decreased with increasing water depth. Below a certain level in any region, called the calcium carbonate compensation depth (CCD), carbonate in essence disappears. The short cores collected by the German Meteor expedition had indicated a lowered calcium carbonate content in the lower part of some Atlantic cores. These low-carbonate sections were thought to be glacial in age. This conclusion was based on the variations in species distribution of planktic foraminifera as indicators of surface seawater temperatures (Schott, 1935). Thus, there were indications that the carbonate content changed through time in deep-sea cores. The long cores now available from the Albatross expedition would reveal striking cyclic patterns in the carbonate content of Quaternary sediments in all oceans.

G. Arrhenius (1952) made a classic study of the carbonate distribution within the cores of the **East Equatorial Pacific**. He found that the calcium carbonate content varied greatly, with an alternation between carbonate-rich and carbonate-poor layers and overall decreasing concentration back in time. The top layer, the Postglacial or Holocene (stage 1), was found to be characterized by low values of calcium carbonate. Below the carbonate-poor surface sediment a carbonate-rich layer (stage 2) was present in all cores, indicating an increased rate of accumulation of carbonate with several subordinate maxima (Fig.15). Arrhenius now sent for dates based on the newly developed C-14 dating method. The radiocarbon datings suggested that the transition between stages 1 and 2 could be approx. 17,000 years in age (Arrhenius, Kjellberg and Libby, 1951). Stage 2 extends back to approx. 70,000 BP according to his titanium chronology. Arrhenius was able to recognize at least 18 carbonate stages, covering the entire Pleistocene and the uppermost Pliocene as well (in Core 62).

Kuenen (1950) had proposed a hypothesis that the lutite sedimentation in the eupelagic environment would be practically constant. Arrhenius made use of this lutite veil concept assuming that titanium content could be used as a measure of the lutite concentration. He developed a titanium chronology, calibrated by the ^{14}C-datings, and used this dating for most of his cores. Kullenberg (1953) criticized his approach, suggesting that titanium accumulation rates could vary significantly. I found that certain older index horizons of the

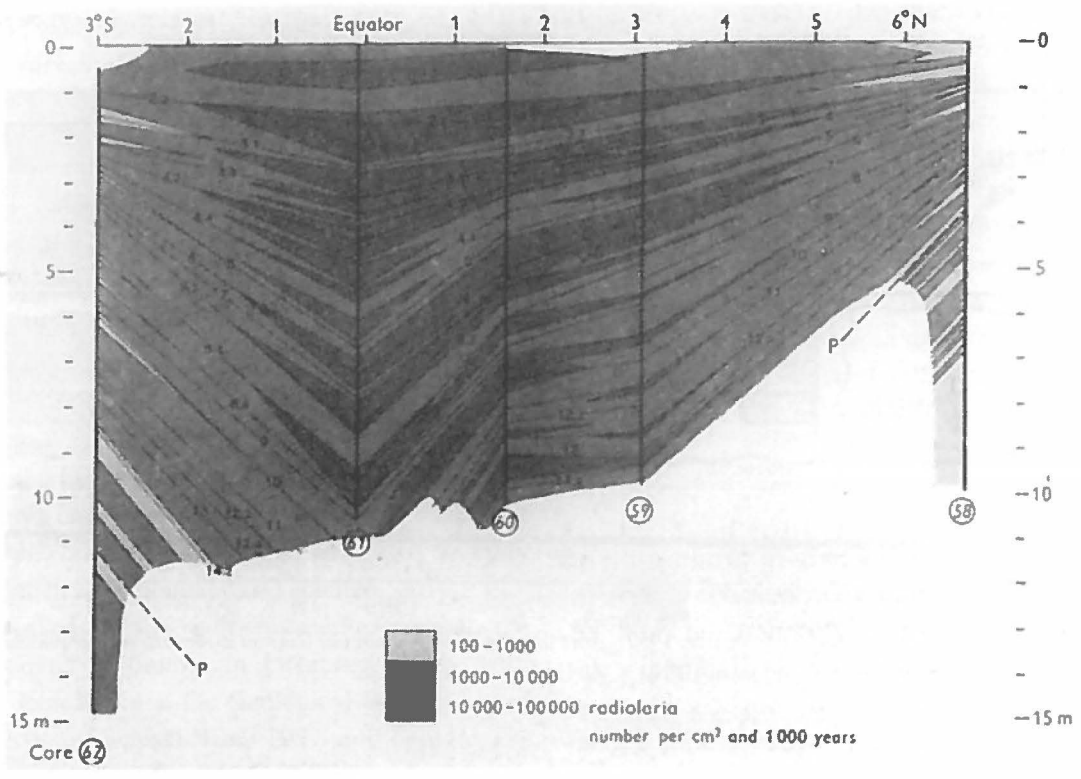


15. Carbonate curves of cores 45 and 60 and their stratigraphic interpretation (from Arrhenius, 1952, 1988). Arrhenius´ carbonate stage 3.3 corresponds to the isotopic stage 5 e (= the Eemian).

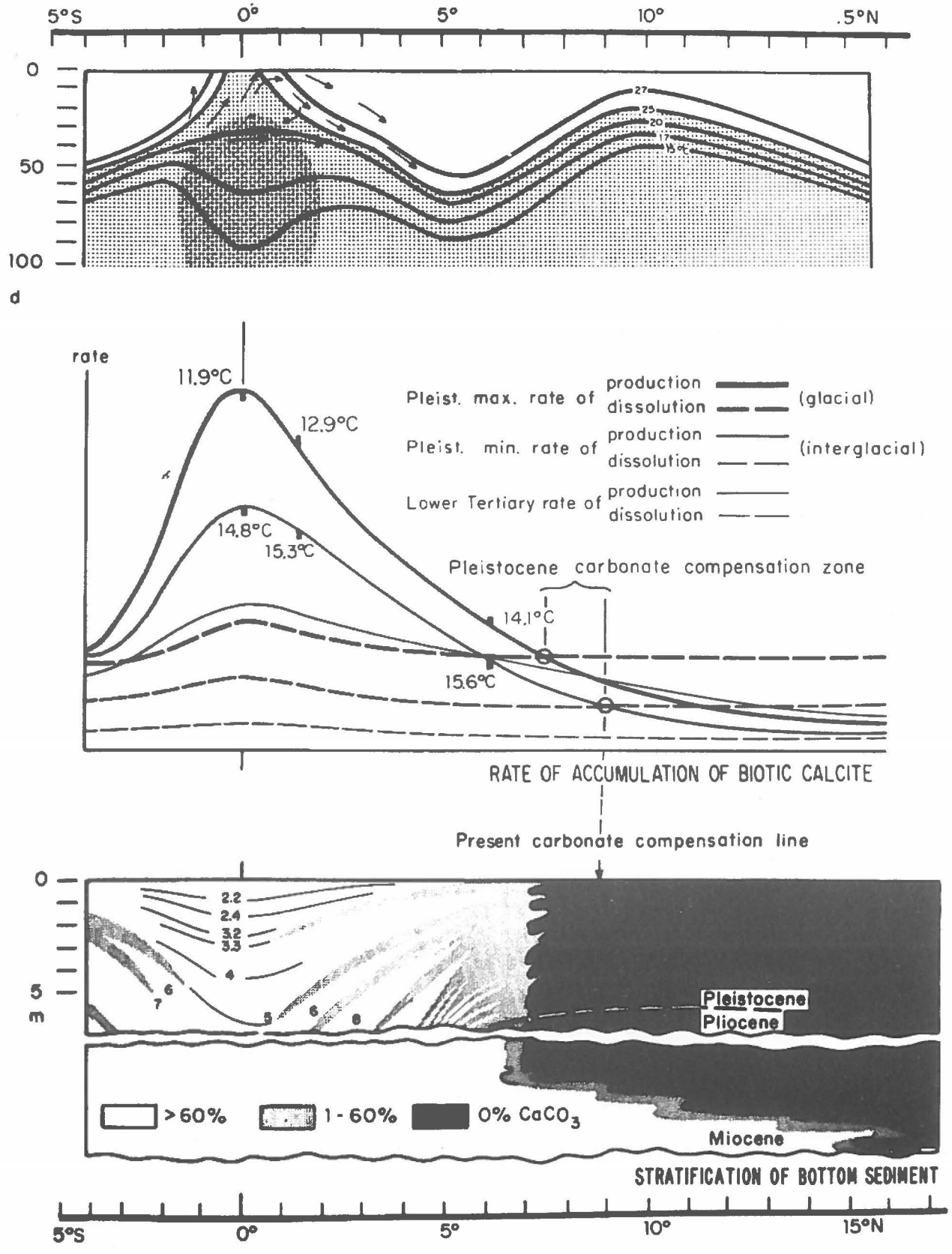
cores 58 and 62 , which I took to be synchronous, differed in age by up to half a million years (Olausson 1961 b). We know now that the lutite veil concept is incorrect: the lutite concentration in the sea water cannot be considered as constant in time. Thus, the titantium chronology must be considered as unreliable. However, it gave the first estimates for long-term accumulation rates which, were, on the whole, of the right magnitude.

Was the high glacial carbonate accumulation due to an increased sedimentation of calcareous matter or to a decreased rate of dissolution of carbonate at the bottom?

Arrhenius (1952, 1988) found that the number of radiolaria per cm^2 and 1,000 years was always enlarged by a factor of two to ten in the high-carbonate stages when compared with the intervening low-carbonate layers (Fig. 16). No signs of corrosion and decomposition of radiolarian skeletons were seen except near the calcium carbonate compensation surface. Also, the content of phosphate suggested an increased rate of accumulation by a factor seven to eight in the last two, most pronounced carbonate maxima, compared to the intervening minima. The centric diatom *Coscinodiscus nodulifer (= Azpeitia nodilifera)* showed interesting features. There were two types of frustules, one with a size maximum around 40 µm, another with a maximum between 80–120 µm, particu-



16. Rate of accumulation of radiolaria > 150 µm in the East Equatorial eupelagic section (from Arrhenius, 1952).



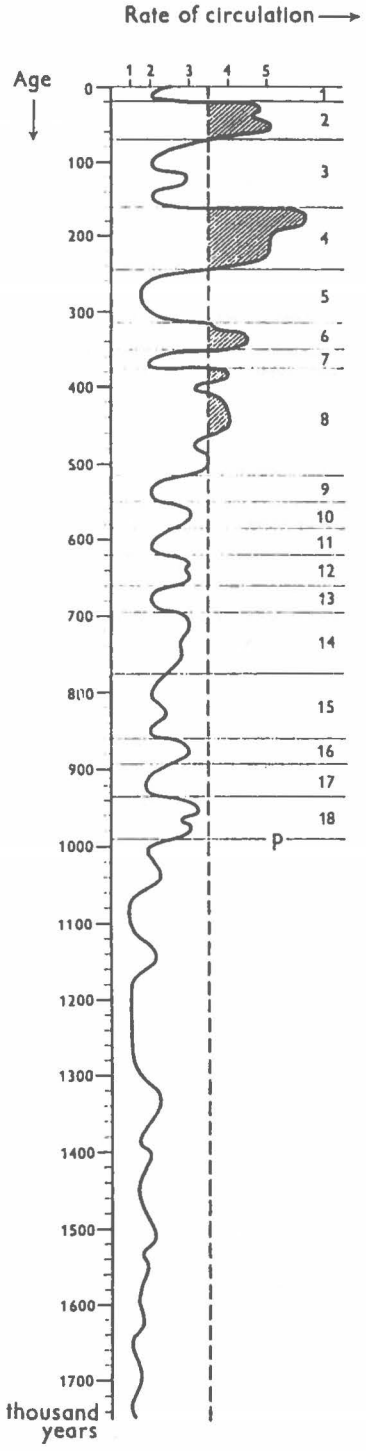
17. Productivity and currents in the Pacific (Arrhenius, 1952). Note that the South Equatorial Current is located at the equator, the Counter Current at 5°N and the North Equatorial Current at 10°N.

larly up from the Saalian glaciation (= from isotopic stage 10) but noticeably further down, at least down to the early cool phase of the early Holsteinian (isotopic stage 14). The large types are common in the glacial stages but have a very low abundance in the interglacial ones. The small size diatom type is frequent throughout the entire Pleistocene, but relatively less abundant in the glacial carbonate peaks.

From this findings, Arrhenius concluded that the high calcium carbonate stages in the Central Pacific are the result of intensified organic production during glacial ages. Berger and collaborators have proposed in a series of publications from 1971 that the effects from Pleistocene productivity changes in the Equatorial Pacific have been negligible in comparison with effects from (presumed) periodic cycles in carbonate dissolution (Fig. 17–18). As will be discussed further down, in my opinion Arrhenius was right in postulating large changes in the rate of organic production in the Indo-Pacific Oceans while the dissolution process at the Indo-Pacific floors has been rather constant during the Quaternary times, contrary to Berger´s proposition.

G. Arrhenius finished his thesis about the Eastern Pacific cores (38–62) in 1952 and subsequently went to Scripps Institution of Oceanography in La Jolla. The studies of the other cores (21–36 and 69–299) were not carried out as planned originally. Most of the necessary investigations remained undone, and only selected studies were performed until 1956. In June1956 I was asked by Prof. Kullenberg to carry out a general survey of the cores 69 –299, to present systematically their composition, stratigraphy and chemistry in both text and figures. After some hesitation for such a large task (and because of the poorly defind conditions of employment), I accepted the offer. I started the work on October 1, 1956. My experience with deep-sea sediments up to that day was practically nil. My graduate thesis (for Fil. lic. examen, about equal to American Ph. D. degree) had dealt with the development and the fossil content in a Holocene mire complex in Western Sweden (Olausson 1957). I was, thus, a trained micropaleontologist, familiar with pollen, diatoms and plant remains, particularly those of the family *Sphagnum.* Besides, I had started to study more chemistry, particularly soil science.

I worked partly with my colleagues at the Oceanographic Institutute in Göteborg, partly with professor Erik Norin at the Geological Department of the University of Uppsala from 1957 until I got my Ph. D. degree in April 1961 when I left Sweden for Miami.



18. Rate of low latitude atmospheric circulation in the equatorial Pacific as a function of time (Arrhenius, 1952). The time-scale is now revised, particularly for the lower part of the sequence. Stage 3 = the Eemian-Odderade interval (≈ 120–60 ka), 8/9 ≈ the Brunhes/Matuyama boundary (≈ 800 ka), 13/14? ≈ Jaramillo (≈ 1 Ma), and P = the Pliocene/Pleistocene boundary (according to Arrhenius), probably the Reuverian/Pretiglian transition, approx. 1.3–1.4 Ma.

After having read Arrhenius´ aforementioned thesis, parts of the textbooks "The Oceans" by Sverdrup et al. (1942) and Kuenen´s "Marine Geology" (1950) during a week or so, I started to work out a classification scheme of the deep sea deposits which I could use. I developed schematic drawings of the compositions, colours and stratigraphy of the cores, and diagrams for presenting their chemistry, and if known, also their grain sizes and fossils content (Olausson 1960 a–d). In the schematic drawings of cores every major constituent was indicated by a sign. Combinations of two or more signs illustrate partly the constituents of the sediment, partly the name of the deposits (Olausson 1960 a–d). My earlier experience of presenting data from peat bogs was here of great use even considering the fact that the depositional environment and the fossil content were quite different. The many discussions during those early weeks in the study with Dr. F. Koczy who soon would leave for Miami are gratefully remembered.

My classification system of the deep sea deposits and the graphical methods to present the core data worked out in October 1956 were subsequently taken over, with modifications, by the scientists of the Deep Sea Drilling Project, in the late 60´s and early 70´s.

The cores from the **Western Pacific** (No. 69–122) can be divided into four groups:

a. The cores 74, 75, 91–98, 104 and 105 have a similar carbonate stratigraphy as those from the Eastern Pacific. The cores 74 and 75 were dealt with by me in two early papers (Olausson 1960 a, 1961 a). Core 74 covered the Pleistocene down to approximately isotopic stage 30 (probably the Eburonian ice age) and core 75 down to around the stage 20 (Elsterian ice age). – The uppermost 8 m of core 91 gives clear indications of about ten ice ages. The sediment below 838 cm is coccolith ooze, in the same way as e.g. core 58 (Olausson 1966), probably of late Pliocene age. – The cores 92–97 consist of Pleistocene carbonate sequences of various length comparable with those from the adjacent stratigraphically wellstudied Ontong – Java Plateau (see e.g. Ocean Drilling Program (ODP) Leg 130, 1993). – Core 104, from the Caroline Basin, consists of chalk ooze composing the Saalian up to the Holocene (= isotopic stages 10–1). Judging by the chemistry diagrams it is likely that the core 105 contains a complete sequence from ca. isotopic stage 26 (Menapian) up to the late Weichselian (stage 2). Core 106 also displays a well-preserved record from the middle – late parts of the Pleistocene.

b. The cores 83–90 are from an area below the present CCD and from a rather productive zone, at and close to the equatorial divergences. Radiolarian oozes are formed here. The cores 85, 88 and 89 were collected from larger depths than 83, 86 and 87 and the tops of the cores consist of red clay with some siliceous fossils remains, indicating severe dissolution there. Core 90 contains 194 cm of radiolarian ooze overlying chalk ooze of Miocene age.

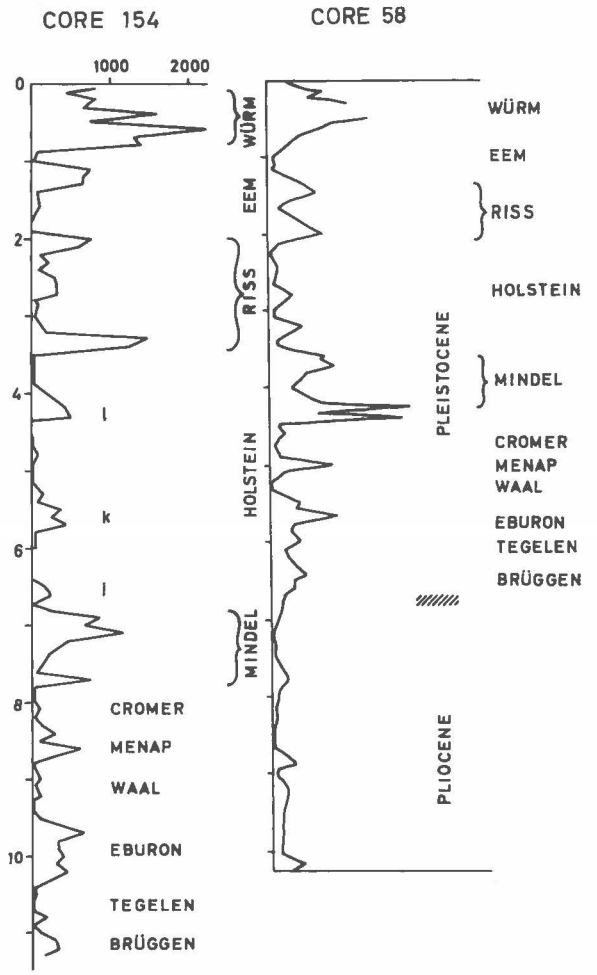
c. The cores 69–73, 77–81, 85, 88 and 89 consist of red clay. Some of these cores shows a rather thin layer of probably Quaternary red clay which rests disconformably on middle or late Tertiary sediments (Riedel 1957, Olausson 1961 a). These complex sequences will be dealt with further below in connection with the pre-Quaternary core sequences.

As can be seen from Fig 20 the area below the Equatorial Counter Current in the Pacific constitutes a transition zone between the high carbonate facies (below the Equatorial Divergence) and the low or non-carbonate facies to the north. It seems that the changes in the productivity are stronger in the South Equatorial Divergence zone than the North Equatorial Divergence zone. This observation has been used to find the cause of the increased supply of nutrients to the glacial Indo-Pacific oceans (see further below).

d. The cores 107–122 are from the Indonesian region. They consist of volcanic material of various grain sizes. Their stratigraphy has not been established.

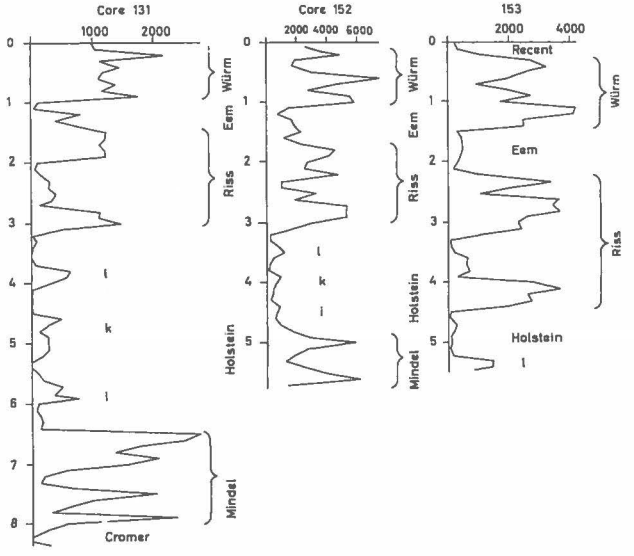
In the **Indian Ocean** the cores 125–158 were raised, 125–133 from a latitude near 10° S, 134–154 from a equatorial zone and 155–158 from a transect in the Somali Basin, 5–10° N. They are described in Olausson (1961 b) and stratigraphically discussed particularly in Olausson (1971a) and Olausson et al. (1971). In the Indian Ocean the South Equatorial Divergence occurs at about 8°S whereas a convergence is found at about 4°S. During the northern summer, a North Equatorial Divergence exists at about 0°. When looking at the map (Fig. 10) and considering the coring depths, one might suppose that cores (128), 131, 152, 153, (154) would be likely to show a correlation between glacials and high carbonate facies. Cores 131 and 152–154 clearly show the same characteristics as cores 58–62, 74–75 from the Pacific, and core 128 has carbonate beds alternating with siliceous ones. Furthermore, radiolarian ooze occurs in the deeper region below the eastern sector of the South Equatorial Divergence (cores 127–133). Below the North Equato-

rial Divergence red clay ought to have been deposited if the carbonate had been dissolved (Olausson 1966, Fig. 22; 0lausson et al., 1971, Fig 2). This suggests that the rate of production was higher below the South Equatorial Divergence (SED) than below the North Equatorial Divergence (NED) in both the Indian and Pacific Oceans during ice ages. This observation of higher productivity below SED than NED is acribed to the present drainage system of the continents and the vertical circulation pattern of the oceans (Olausson 1967; Fig. 19).



19. Stratigraphy of Core 154 from the Indian Ocean and the Pacific core 58. The curves give the number of entire foraminifers per g sediment. Due to higher rate of production during glacials and a rather constant supply of bottom water to these two oceans, the number of preserved tests increase considerably during the ice ages, particularly the more intense as the Mindel (= Elsterian or the isotopic stages 20–16), Riss (= Saalian or the isotopic stages 10–6), and Würm ice ages (= Weichselian or the isotopic stages 5d–2). The other stratigraphical names are explained in chapter VI:1.
The rate of burial of tests reflects the export of nutrients from the North Atlantic to the Indo-Pacific Ocean, which according to my nutrient transport theory, is climatically controlled.
My $\partial^{18}0$ and $\partial^{13}C$ of Core 154 are published by Naidu et al., (1993).

The other Indian Ocean cores collected by the Albatross Expedition, away from the eqquatorial zone, can be divided into three groups, in which the correlations between glacials and high carbonate facies is less obvious. These are Cores 155–158 from sites under the influence of upwelling off the Somalian coast, Cores 140–144 and Cores 145–149. The curve obtained from analyzes of the carbonate fraction of core 131 exhibits the same trend as the Pacific curves. It extend down to the isotopic stage 20 (Cromerian). *Globigerina inflata* reaches a characteristic maximum percentage in the isotopic stage 16 (Upper Elsterian) in this core as well as in other Indian Ocean cores. Core 152 extends approximaterly down to the ca isotopic stage 20 (22) (= Cromer/ Elsterian transition). Core 153 penetrates into the isotopic stage 12 (=Holstenian stage), and Core 154 consists of the whole of the Pleistocene (Fig. 19–21). All these cores have carbonate maxima in their glacial stages (Olausson 1971 a). The productivity changes together with differences in the dissolution of carbonates with increasing depths, may be the explanation of the shifts from siliceous facies at Core 133 via red clay facies at core 134 to carbonate facies at core 137. A turbidite present between 154–392 cm of core 137 makes that core difficult to interpret stratigraphically. The cores 139–140 are from the distal part of the Ganges submarine cone (fan). The cores 141–143 are both late Pleistocene in age but their stratigraphy has not been studied in detail. The uppermost 4 m of core 144 consists of (middle-)late Pleistocene sequences



20. The content of entire foraminiferal tests relative to depths (in meters) in the three Indian Ocean cores 131, 152 and 153, and their time-stratigraphic interpretation (as in Fig. 19)

resting on two beds of different chalk oozes. Core 145 consists of three beds where the upper c. 6.35 m is Pleistocene chalk ooze, followed by 1.2 m of abnormal foraminiferal sand, below which normal carbonate deposits are found. The cores 145–149 have a complicated, and disturbed stratigraphy and were therefore not studied in detail (see further Olausson 1960 b). – The cores 155–158 are discussed further below in connection with reconstruction of the past monsoon regime, Chapter X (see also Olausson et al. 1971 b).

The **Red Sea** and **Mediterranean** are protected from cold, aggressive bottom currents by shallow sills: the Bab el Mandeb is about 145 m deep and the maximum water depth in the Strait of Gibraltar is 394 m. Since the carbonate content in deep-sea sediment from the open oceans generally is constant down to at least 2,000 m water depth, variation in the carbonate content in the Red Sea and Mediterranean must be due to dilution by non-carbonate matter. In the Mediterranean we cannot use methods based on variation in the dissolution capacity of deep waters as applied in e.g. the North Atlantic. The fundamental difference in preservation patterns in the two regions makes the transfer of the basic foraminiferal (CLIMAP) data from the recent North Atlantic to the glacial Mediterranean (Thiede, 1978) highly problematic.

The cores 163–169 are from the Red Sea and cores 176–211 from the Mediterranean.

The **Red Sea** cores are all highly calcareous. Due to the alkaline bottom environment pteropods also occur in these cores, mainly in their uppermost part. The aragonitic pteropods are generally easily dissolved; we found these fossils in open ocean sediments only in a thin layer from the Weichselian/Holocene transition off western Africa, as outlined in the discussion about the preservation spike further below. In the upper part of the Red Sea cores aragonitic materials are included in concretions, a remnant of a highly saline or evaporating phase of the Red Sea during the glacial maximum. In a special chapter our Red Sea cores and their evidence about past changes in the hydrography of the Red Sea basin are discussed.

The calcium carbonate content of the eastern **Mediterranean** cores is generally rather low in the eastern Levant, particularly off the Nile delta and in the southwestern Mediterreanean where eolian supply is pronounced. Elsewhere, in the pelagic area, the content is generally 30–40 per cent, somewhat less off the volcanic areas near Italy. There is no obvious relation between the carbonate content and climate.

The 50–100 cm long **Atlantic** cores collected by the German Meteor Expedition revealed a marked decrease in the carbonate content in their lowermost part, interpreted as glacial in age. The downcore decrease in carbonates was discussed both by C.W. Correns and by W. Schott. Schott (1935) assumed that during glacial epochs surface water in the tropics was cooled by as much as 10° C so that the extraction of calcium carbonate was much reduced.

The Albatross cores, collected in the North Atlantic, all showed high carbonate content in the upper half meter or so and then low and high carbonate facies alternating downcore (Olausson 1960 d). ^{14}C-dating of the topmost carbonate maximum showed that the maximum is Holocene in age and that the uppermost minimum is about 18,000 years in age. The increase of the carbonate content occurred during the last deglaciation. This general carbonate stratigraphy has the opposite trend as that in the Indo-Pacific Ocean. Thus, a carbonate minimum in the Indo-Pacific corresponds in time with a carbonate maximum in the Atlantic, and a carbonate minimum in the Atlantic, particularly in the North Atlantic, is synchronous with a carbonate maximum in the other two oceans. That pattern was a new discovery. What is the explanation of this pattern of opposite carbonate alternations? That was the problem I started to investigate during the Christmas holidays of 1960.

3. *The cause of the calcium carbonate variations*

A. General comments. Half a century ago the calcium carbonate-dissolving capacity of ocean deep and bottom waters during the Quaternary was ascribed to the low temperature and high content of carbon dioxide in abyssal waters set free by the decomposition of organic matter. This capacity was further associated with Arctic bottom current and, especially, Antarctic bottom current. Solution was known to be more active with increasing depth. Below a sea floor depth of c. 4,000 m there is a rapid decrease in the carbonate content, and 500–1,000 m further down all carbonates are dissolved in the oceans. The level in the ocean where the supply of carbonates is equal to the rate of dissolution in the bottom water is called the **calcium carbonate compensation depth, CCD** for short.

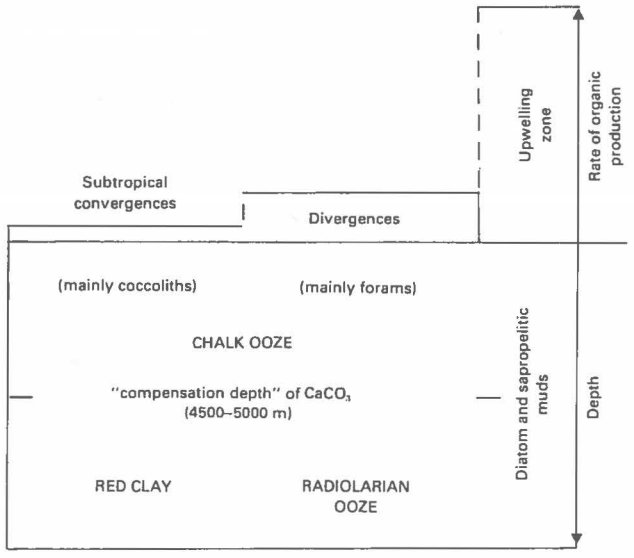
It was thought earlier that during the Tertiary when the bottom water was assumed to have been

much warmer than at present and when no driving power for the bottom currents in the form of high-latitude sink water formation from cooling was thought to have existed, no extensive layers of red clay could have been formed. However, some of the Albatross cores in the Pacific which reached down into Tertiary did also consist of red clay.

Thus, at least this basic assumption about the calcium carbonate accumulation seemed incorrect. Generally speaking, also at lower water temperature an equilibrium between the bottom water and the falling carbonate matter could be established if the carbonate-dissolution capacity (= acidity) is/was not too large compared with the supply of biogenic carbonate from the surface layers (as at present high latitudes).

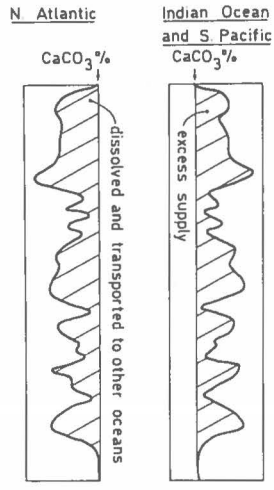
During 1959–60 I had started to print my basic reports of the cores from the three oceans and the two adjacent seas, the Mediterranean and the Red Sea as well (Olausson 1960 a–d). I had also begun to analyze and interpret the data from the Eastern Mediterranean which were to be the main part of my Ph.D. thesis. During that time in December 1960, as mentioned before, I began to theorize about the distribution pattern of the carbonate oozes on the sea floor and its connection with the depths and the surface current systems. Furthermore, I found that coccolithophorids were the most important constituent in carbonate deposition in low-productivity (oligotrophic) areas such as the anticyclonic gyres. The number of foraminifers typically increased towards high-productivity regions, that is divergences and other upwelling areas. I could thus tell which type of deep-sea deposit we could expect to find below each hydrographic province when taking into account the depth of the sea floor as well.

Fig. 21 is meant to illustrate the relationship between the quantitative production of planktic foraminifers and coccolithophorids. We could get a high content of calcium carbonate in low-productivity regions due to low supply of siliceous organisms and low supply of inorganic matter. We could also get high content of calcium carbonate in high-productivity areas such as below equatorial divergences. An increased rate of organic production would result in a higher content of foraminiferal tests compared with periods of low productivity. These results suggested that the carbonate content could be used for stratigraphical purposes in oceans where the cause of a changing carbonate content is a varying intensity in the dissolution processes at the sea floor. However, in areas with more or less constant dissolving capacity of the bottom water but with a varying productivity the number of foraminiferal tests per unit of time, or in case the rate of accumulation is not known, number of (entire) foraminiferal tests > 150 μ per gram or per cm^3 could be used.



21. The distribution of deep-sea deposits in relation to water depth and the rate of organic production. The supply of detrital matter, larger than the clay fraction, is supposed to be small.

As stated above, Arrhenius found that a low carbonate content is the characteristic facies for the present central Pacific. I found a similar trend in the central Indian Ocean, but the opposite trend for the North Atlantic (Fig. 22). In the paper which I wrote during

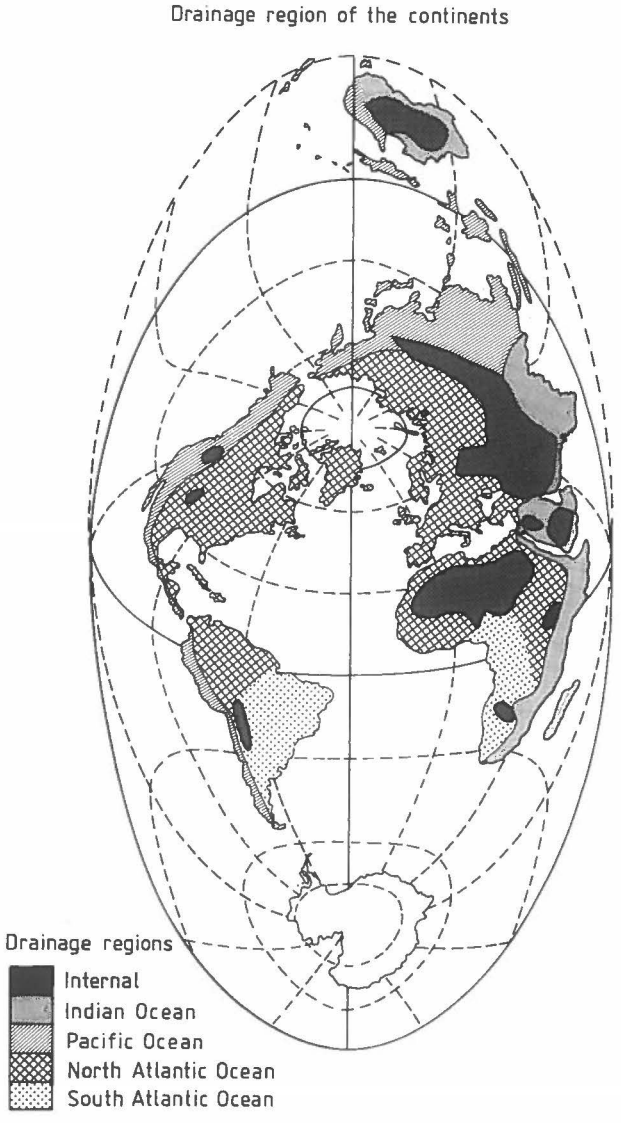


22. If the vertical circulation of the oceans was constant in time the accumulation of carbonate could be constant in each ocean. The distribution of $CaCO_3$ would then only be straight lines as those below the arrows at the diagram. However, any change in sinkwater formation will alter the supply of dissolving matter to the bottom (CO_2 and organic matter), causing variation in dissolution at the North Atlantic floor, and as a consequence, a reverse variation in the Indo-Pacific Ocean.

Christmas of 1960 I accepted Arrhenius´ suggestion that an increased rate of the atmospheric low-latitude circulation was the cause of the increased carbonate deposition during glacial ages. Since the prevailing winds govern the strength of surface currents, they could, if increased, be responsible for enhanced upwelling in the equatorial divergences during glacial times (Olausson 1961 b, p. 17).

During my further studies of Indian Ocean and Atlantic cores I understood that the present carbonate distribution mirrors the drainage system of the continents and that the vertical circulation of the oceans changed this pattern, particularly during glacials (Olausson 1967). An increased deposition of carbonate in any one place is also a geoeconomical question. The accumulation at the sea floor of each element cannot be larger than the supply; excess accumulation of an element or a compound such as calcium carbonate could only be possible during short periods. Furthermore, the cores from the Somali basin suggested a weakening of the trade winds during northern summers during the peak of the last glacial (Olausson et al., 1971). After some centuries of tropical coolings this was followed by a gradually intensified low latitude circulation. The present position of the eastern position of the Somali Current was reached 5,000 to 8,000 years after the climatic deteroriation. The flow during the northern winters was kept low during the whole cool wave (see further Chapter X, and Fig. 61). Already Flohn (1952, 271) had pointed out that the wind velocity at low latitudes (5° S to 5° N) must be rather constant "da hier der Luftdrucksgradient keine Rolle spielt und die verschwindende Coriolis-Kraft alle aufkommenden Stürme sehr rasch wieder abdämpft" (cf. maps of the tracks of the tropical cyclons, e.g. in T. Bergeron, Quart. Journal, 1953). The suggestion by Arrhenius of an increased trade-wind induced transport of equatorial surface water during glacial times was thus found to be incorrect. The role of a varying transport of calcium carbonate, silica and nutrients from the Atlantic to the other oceans as a cause of the variation in the accumulation of carbonate and other biogenous sediments became more clear to me during the mid 60´s (Olausson 1965, 1967). Later, in the invited talk and the manuscript delivered at the SCOR´s international symposium on the Micropaleontology of Marine Bottom Sediments (held at Cambridge 10–17 Sept. 1967) I developed my nutrient transport theory in more detail (Olausson, 1971 a).

The land area drained into the Atlantic is nearly twice as large as that draining into the Indian and the Pacific Oceans taken together (Fig. 23). Therefore, the Atlantic, particularly the North Atlantic, receives much more weathering products from river discharge than the other oceans. This illustrated by the river discharge of calcium and silica to each ocean and applies probably also to most other elements derived from continental weathering.



23. The drainage region of the continents. Note that about two-thirds are drained into the Atlantic Ocean. Therefore, the Atlantic, particularly the North Atlantic, receives much more weathered products from rivers than the other oceans do. However, dissolution etc. combined with water exhanges between the oceans, smooth out more or less effectively the differences caused by the present drainage patterns, the more the higher the intensity of the vertical circulation is (cfr. Fig. 25).

Table 1: River discharge of dissolved calcium and silica; in 10^{13}g/y, (Olausson 1967):

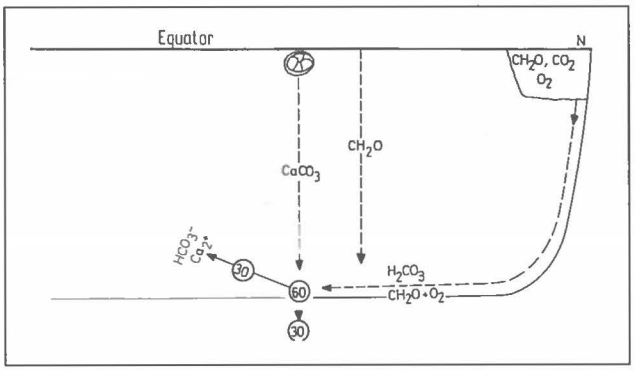
	Ca	SiO_2
Atlantic Ocean	35.9 (78 %)	24.5 (57 %)
North Atlantic	32.2 (70 %)	17.2 (40 %)
South Atlantic	3.7 (8 %)	7.3 (17 %)
Indian Ocean	4.6 (10 %)	14.2 (33 %)
Pacific Ocean	5.5 (12 %)	4.3 (10 %)
	46.0	43.0

During the last 20 years or so an intense hydrothermal flow through the oceanic crust has been discovered. Annually some 170 km^3 of sea water circulates through the uppermost part of the ocean floor. The hydrothermal waters return to the oceans at or very close to the spreading axis system (submarine rift), delivering many elements and chemical compounds to the oceans from magmatic sources. We now know that hydrothermal activity must occur throughout the 50,000 km length of the spreading axis system, wherever a magma chamber develops. For calcium, the supply from hydrothermal activity is estimated at $0.73 \cdot 10^{14}$ g/y and for SiO_2 at $0.82 \cdot 10^{14}$ g/y. If true, the hydrothermal supply of Ca constitutes ca. 15 per cent of the total supply and for silica 41 per cent. If we assume as a first approximation, that the total supply of these elements to each ocean is related to the length of the midoceanic ridge in each ocean, some 25 per cent of each element from the hydrothermal vents should be discharged into North Atlantic, 15 % to the South Atlantic, 40 % to the Pacific and 20 per cent to the Indian Ocean. A combination of these two data gives the following approximate percentages of the (yearly) distribution of Ca and SiO_2.

	Ca	$Si0_2$
North Atlantic	64 %	38 %
South Atlantic	9	14
Indian Ocean	11	29
Pacific Ocean	16	19

These percentages are close to those given on the basis of the river discharge only (Olausson 1967, p. 246, and 1971, p. 376), thus, there is no need for revision of my theory, from the findings regarding hydrothermal activity.

As we know, there is no surface outflow from the North Atlantic ocean. The bulk of the net out-transport of c. $1.8 \cdot 10^{14}$ g/y of calcium occurs in the cold-water sphere. The calcium is removed from the surface water by organisms, chiefly foraminifera and coccolithophorids and, to a minor extent, from the bottom water by benthic foraminifera (Fig. 24). The settling time compared to the time during which the calcareous remains are in contact with the bottom



24. The vertical circulation of the North Atlantic – Nordic Seas. On average about sixty foraminifers settle to the bottom per square centimeter per year. Theoretically, thirty of those are presently dissolved and the dissolved substances are transported to other oceans.

water is neglegible. Hence, the solution of carbonates can be supposed to occur chiefly at the bottom (recycling due to predation in the warm-water sphere is here discounted). Solution here can be described by the following two equations:

$$CaCO_3 + H_2CO_3 + CO_2 \rightarrow Ca^{2+} + 2\,HCO^-_3 \quad (1)$$
$$CaCO_3 + CH_2O + O_2 \rightarrow Ca^{2+} + 2\,HCO^-_3 \quad (2)$$

where CH_20 stands for organic matter (soft organic remains in various degrees of decomposition). Large, soft organic remains (as pellets, aggregates) may reach the bottom due to their own weight, but smaller organic particles, colloidal organic matter, and organic matter in true solution seem to reach greater depths chiefly by descending water masses. In a litre of the water that sinks down to considerable depths or to the bottom in the northernmost Atlantic there may be some 0.5 mg of carbon, 5–8 ml of oxygen and some 5 ml of free carbon dioxide. It was calculated earlier (Olausson 1967) that at present about 70 per cent (here above 64 %) of all carbonates are discharged into the North Atlantic and that 35 per cent of the river-transported Ca is accumulated as carbonates at the North Atlantic floor while the other part is dissolved, exported and deposited elsewhere. Theoretically 3 milligram $CaCO_3$ is deposited in the North Atlantic per square centimeter per year on average. With the average weight of a test of an adult foraminifera being 0.05 milligram (generally between 0.1–0.01 milligram), about sixty tests settle to the bottom per

square centimeter per year, and thirty of them are dissolved (Fig. 24). In order to dissolve all these tests on the North Atlantic floor all the free carbon dioxide in about $2 \cdot 10^4$ km^3 of sink water is needed (eq. 1), or all carbon in 10^5 km^3 sink water (eq. 2). The present production of sink water in the northernmost Atlantic is estimated at about 20 Sverdrups (20×10^6 m^3 per sec.) or 6×10^5 km^3/y. During its flux southward – in the direction of Pacific – organic particles settle down, adding more organic matter to the bottom (Fig. 25 a). Therefore it seems possible that the processes described in eq. 2 may be the most important, particularly at lower latitudes.

An attempt to test this transport theory was made by calculating of the Ca transport across the equator from the alkalinity (Olausson 1971 a, Fig. 29.3). Later on I measured the Ca directly from a profile there. I found that the concentration of Ca was 3 mg/l less than in normal sea water of the same salinity down to approx. 500 m. I found higher Ca/Cl ratios at a water depth around 1,000 m (where Mediterranean water occurs) and close to the bottom. In both cases Ca was 2 mg/l higher than in normal sea water. Higher Ca/Cl ratio was also measured off the Antarctic at a depth of about 1000 m.

From aformentioned equations (1) and (2) and from the discussions above it follows that the carbonate content of sediments, the solution rate at the bottom, as well as the pH of the environment, depend on the supply (in moles per unit of time) of calcium carbonate (chiefly the vertical supply), and organic matter (both vertical and lateral supply), carbon dioxide and oxygen (both by descending water mass). In areas flushed by North Atlantic sink water ("North Atlantic Deep Water") oxygen is present in excess and if we assume the proportion of calcium carbonate and organic matter settling down from the overlying surface water to be rather constant; changes in the solution of calcium carbonates and thus also changes in the calcium carbonate of the final sediment are related to the intensity of the vertical circulation.

Changes in the accumulation rates of calcareous matter can be recognized by carbonate analyses of sediments. However, variations in percent $CaCO_3$ can not only be due to varying intensity in the dissolution at the bottom but also be caused by changes in the supply of non-calcareous matter. Therefore, the percentage of calcium carbonate can be used as measure of paleoproductivity or as a proxy for dissolution only where solution and/or production have changed considerably. The first step in the solution of foraminiferal tests is fragmentation. As stated above, I found fragmentation or the content of entire tests to be a more useful methodthan measuring the percental distribution of carbonates. This proxy for preservation also can be used in areas where the carbonate content proved to be rather constant. Small changes in molar ratio: $CaCO_3/(CH_2O + CO_2)$ at the bottom can be studied by the degree of fragmentation. Essentially the amount of particles > 65μ or > 150 μ could also show the same, but I have found that the number of entire tests was a more useful method. – In the chapter about foraminiferal analysis I will discuss the problem with selective dissolution, how changes in the dissolution may change the proportion between species of different resistance against disintegration.

B. The nutrient transport theory. Organic matter may be more exactly characterized as $(CH_20)_{106}(NH_3)_{16}H_3P0_4$. As the name says, siliceous organisms also contain silica. Oxidation of organic matter will thus not only release carbon dioxide but also compounds containing nitrogen, phosphorus, silica etc. These elements together with calcium are released from the North Atlantic floor and exported to other parts of the world ocean. I suggested (Olausson 1965, 1967) that an ice age (or a cool part of it as that ca 25,000–8,000 B.P.) results in the following changes (see Fig. 25 B):

(i) intensified formation of deep and bottom water in the north Atlantic due to intensified cooling and ice sheet formation on adjacent continents, reduced fresh water supply to the Arctic Ocean – northernmost Atlantic (due to the ice sheet formation and the closure of the Bering Strait) , which

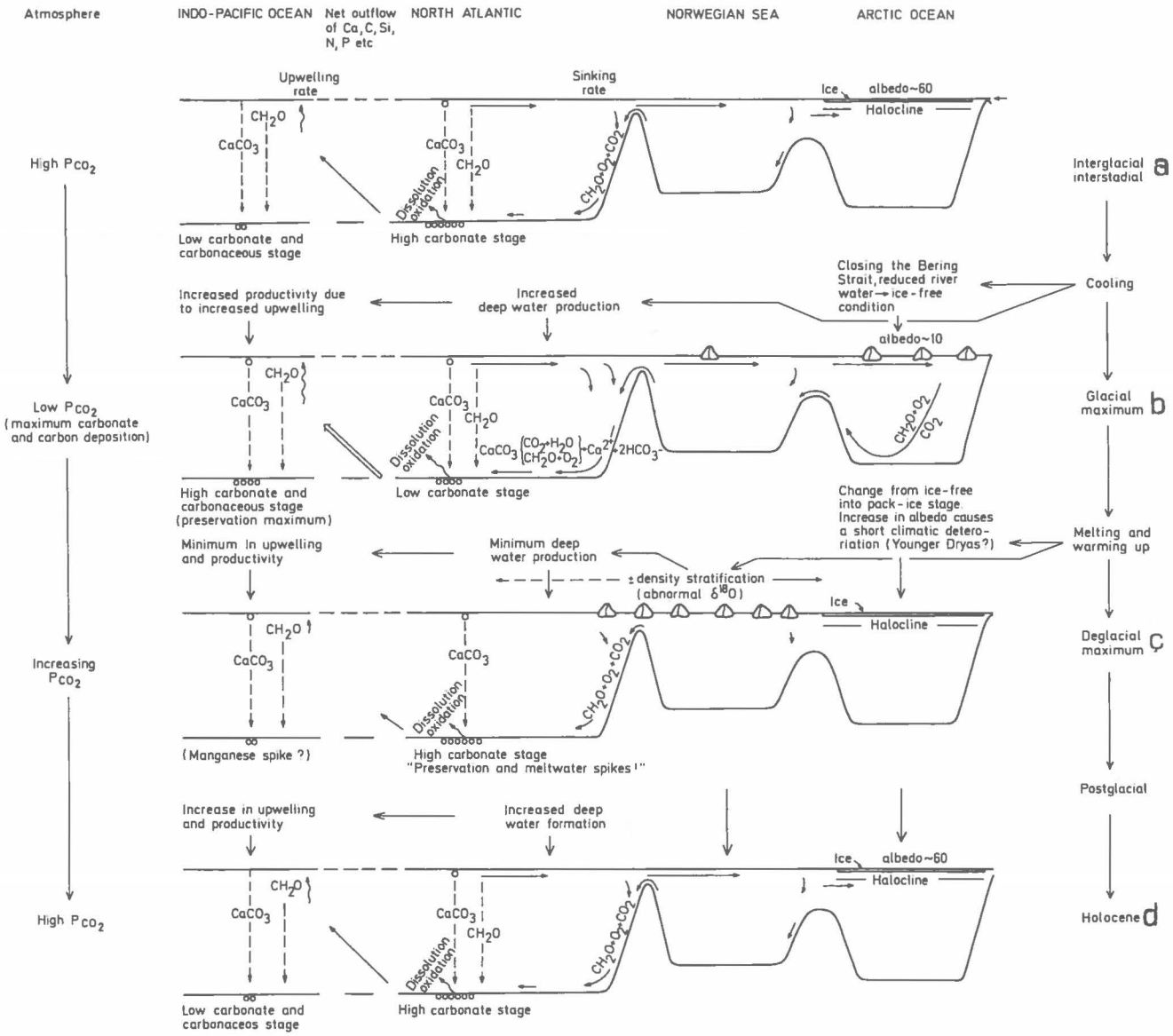
(ii) causes an increase in the supply of carbon dioxide, oxygen, and organic matter to the bottom in the North Atlantic, which

(iii) causes an increase in the oxidation of organic matter (or $C_{106}N_{16}P$) , which

(iv) together with the increased supply of carbon dioxide causes an increase in the solution of calcite, apatite, etc., which

(v) results in an increase in the amount of calcium, phosphorus, silicon, nitrogen etc. that are exported out from the North Atlantic.

(vi) Furthermore,increased sink water formation in the northernmost Atlantic causes an increased overall upwelling in the Indo-Pacific Ocean (the magnitude of the descending water masses must be equal to



25. The abyssal circulation of the oceans during a glacial phase and some processes affected by these exchanges between surface and deep waters (redrawn from Olausson 1969, 1985).

the magnitude of the upwelling waters), which together with (v)

(vii) causes an increase in the rate of organic production in the Indo-Pacific ocean due to increased availability of nutrients there, which

(viii) causes an increase in the rate of accumulation of calcium carbonate and, as a consequence of that, increase in the number of entire foraminifera (Arrhenius, 1952, 1988; Olausson 1961 b, 1965, 1967, 1971, 1985), an increase in the deposition of phosphate, and probably also an increase in the size of *Azpeitia (Coscinodiscus) nodulifera* because of the increased supply of nutrients (Fig. 14–20, 22, 25 B). During deglaciation the situation will be the reversed but in a retarded way. We can use the last deglaciation as a model to illustrate these processes (Fig. 25 C).

The peak of the last glacial epoch occurred around 22,000 B.P. in the Southern Hemisphere, and at 18,000–19,000 B.P. in the Northern Hemisphere (Salinger 1981). From these periods onward there was a gradual increase in temperature starting in the tropics, reaching a maximum in the Australia-New Zealand area at ca. 9,000 B.P. and at ca. 6,000 B.P. in the Northern Hemisphere (l. op. cit.). There is no exact data as to the lowest sea level stand which presumably occurred at 18,000–19,000 B.P. It would be safe to consider -100 ± 20 m as an approximate minimum sea level stand. Most of the ice stored mainly in the form of North American and Eurasian continental ice sheets, was melted before the end of the Preboreal

subages, particularly during the Bölling-Preboreal subages (ca. 13,000–9,000 B.P). About 80 % of the meltwater from the Pleistocene ice sheets was drained into the North Atlantic-Arctic Oceans, ca 5 % into the North Pacific and the rest into the Southern Ocean. Thus, the change in sea level was principally caused by the ice sheets forming and melting in the Northern Hemisphere.

The influence of meltwater on the circulation and chemistry of the ocean probably was first mentioned in Olausson (1961 a). There I argued that the Eastern Mediterranean became stagnant during deglacial times due to meltwater discharge via the Black Sea. Subsequently I discussed the impact of meltwater discharges to the North Atlantic (ca. 20 x 10^6 km^3) during a time of increasing temperature (Olausson, 1965, p. 232). I proposed that sink water formation was reduced during deglacial times (l.op.cit. p. 237), and that the abnormal $\partial^{18}0$ signals found in sediment deposited during the intense deglaciation phases were a result of the meltwater discharge, that is a meltwater contamination, subsequently named "meltwater spike" (Berger, 1978).

The maximum in the meltwater discharge into the North Atlantic occurred within the interval between 13,000 and 9,000 B.P. On the average the global annual rise of the sea level has been of the size of 1 cm. If uniformly spread over the North Atlantic Ocean this amount is equal to a water layer of 6 cm, mixed into the upper 100 m of the area directly influenced by that discharge. This meltwater discharge caused an annual lowering of the surface salinity by 0.04 °/oo, and during more intense deglacial times, such as Alleröd and Preboreal, at least twice that amount (Olausson 1965). Both the temperature rise and the meltwater input may have changed the density of the surface water in the northernmost Atlantic equal to a salinity drop of 1–2 °/oo, thus creating a (quasi)stagnation during intense deglacial period(s) such as Alleröd and Preboreal subages. During the former subage the inflow from the Pacific into the Arctic Ocean started, today contributing 0.1 Sverdrup of freshwater (= the fresh water component in the inflowing low-saline water). This quasi-stagnation in the northern Atlantic (Olausson 1965, 1969, 1985) must have retarded or stopped the sink water formation as long as the density stratification existed the year round. Simultaneously there must have been an equal decrease in the rate of upwelling in the Indo-Pacific Ocean. The following occurred during the quasi-stagnant phase (see Fig. 25 C):

(i) in the Atlantic: (1) decrease or cessation of supply of CO_2, O_2, and organic matter to the sea floor due to reduction of the amount of descending water masses; (2) reduced dissolution and oxidation on the bottom ("preservation spike"), formation of e.g. the pteropod layer. This type of fossil is usually dissolved in cool and aggressive sea water (see Olausson 1969, Berger 1977); (3) reduced volume of the outflowing bottom water and a decreased concentration of its nutrients; (4) increased burial of calcareous and carbonaceous matter in the Atlantic.

(ii) in the Indo-Pacific ocean: (1) decreased upwelling due to the reduced sink water formation in the North Atlantic; (2) decreased production and sedimentation of organic remains and tests; (3) decreased burial of e.g. $CaCO_3$ and organic matter; (4) eventual increase in inorganic precipitation, e.g. manganese due to the very low rate of accumulation (see Berger et al., 1983).

(iii) in the ocean water on a global scale : (1) gradually increased concentration of e.g. $\sum CO_2$ in the deep water; (2) gradual decrease in salinity starting in the North Atlantic.

(iv) in the atmosphere: (1) increase in pCO_2 (see further below and (Fig. 25 C).

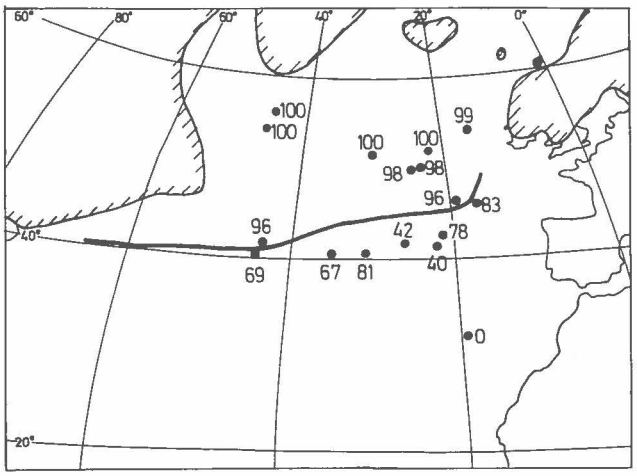
When the meltwater was well mixed into the ocean waters the present type of abyssal circulation was re-established during the time of the Boreal subage and has since prevailed rather unchanged (see Fig. 25 D).

C. The ice-age North Atlantic. One of the main topics of the Albatross expedition was to study the Pleistocene climate. Since the last ice ages affected mainly the polar-subpolar continents of the Northern Hemisphere, it was necessary for the workers of the core material to study particularly the past changes in the hydrography of the northernmost Atlantic.

The cores 297 and 299 from the West European basin (43° and 47°N) contained coarse-grained horizons with iceberg-transported matter. Particularly in the late Weichselian substages of these cores coarse-grained basaltic particles and shards of volcanic glass were frequent (Olausson 1960 d). These layers indicate cool periods when large icebergs drifted into the area.

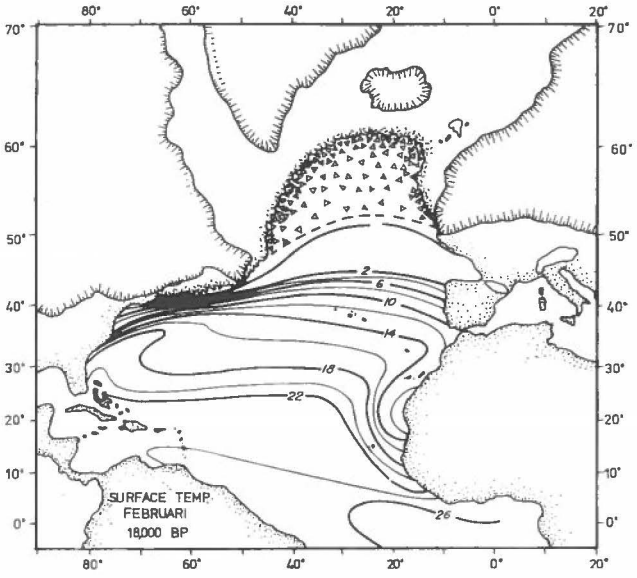
The past location of the Gulf Stream was reconstructed by a method developed by D. B. Ericson

(1964). The coiling direction of the foraminifer *Neogloboquadrina (Globigerina) pachyderma* was found to be to the right in the Gulf Stream – North Atlantic Drift, while this species was left-coiled in the water north of the North Atlantic Drift, i.e. in the cooler Subarctic Water Mass. Within a narrow transition zone the coiling direction changed rapidly, particularly in the western North Atlantic. I used this method for the data from the Albatross cores and showed the southernmost (ice-age) position of the boundary between aforementioned water masses (Fig. 26). It was shown (Olausson & Jonasson 1969) that the boundary had an E-W direction from Cap Cod to 40°W, 40°N, from where it got an ENE direction to a position near 20°W,45°N where the front became unclear or disappeared as it does today. I interpreted this absence of a front as an indication that the North Atlantic Drift also penetrated into the Norwegian Sea during the glacial times as it does today. – Subsequent mapping of this boundary shows the same pattern for the last ice-age maximum but gives more details as to the deglacial times (e.g. by Ruddiman & McIntyre,1981). However, the latter authors found that the Norwegian Sea was outside the reach of the Gulf Stream at that time.



26. The boundary between the Gulf Stream Water Mass and the Subarctic Water ca. 18,000 B.P. The numbers give the percentages of the left-coiled *Neogloboquadrina (Globigerina) pachyderma*. The Polar Front seems to disappear east of 40°E (as today) where the North Atlantic Drift had passed into the Nordic Seas as later on.

Weyl (1968) advanced an interesting hypothesis about the glacial Atlantic. He simply assumed the presence of a fresh-water lid all over subarctic seas in the northern Atlantic during the ice-age. However, he did not explain how this lid could be formed or if on the whole it was theroretically possible. Nonetheless,

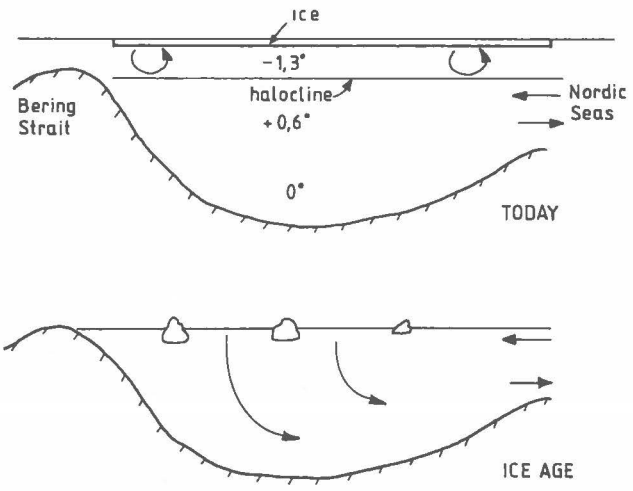


27. Surface-water isotherm map for February 18,000 B.P., in degrees Celcius according to CLIMAP (redrawn from McIntyre et al., 1976)

the freshwater surface layer has then been taken as granted by later authors and the CLIMAP (McIntyre et al. 1976) has in their map covered these adjacent seas with sea ice without any further proof of its existence (Fig. 27).

Perennial sea ice can only be formed on cool, stratified sea water. Therefore the question whether or not a high-latitude sea basin has been ice-covered or not is a salinity problem. Today the **Arctic Ocean** is covered by a layer of sea ice 3–4 m thick sea-ice. The present salinity stratification of the Arctic Ocean is to one half a result of the river influx from the adjacent continents, and to the other half part of the inflow of less saline water from the Pacific via the Bering Strait. Below the halocline, at ca 200 m depth in the Arctic Basin, and down to about 1,000 m the water is comparatively warm (Fig. 28). Also further down the water is still well above the freezing point. The heat content below the halocline is sufficient to melt ice some 20 m thick. Consequently, if this heat could be released upwards, it would be more than sufficient to melt the present ice cover. The halocline, however, prevents this heat exchange.

The sill depth at the Bering Strait is about 40 m, and it is commonly accepted that this strait was a dry area (land bridge) during the last glaciation(s). Thus, the salinity difference between surface and subsurface water in the Arctic Basin was reduced to some 50 % during the onset of a glacial, other factors being equal. The lower part of the Mackenzie and the North Euro-



28. The hydrography of the Arctic Ocean today and ca. 18,000 B.P. Due to lack of fresh-water supply the halocline disappeared.

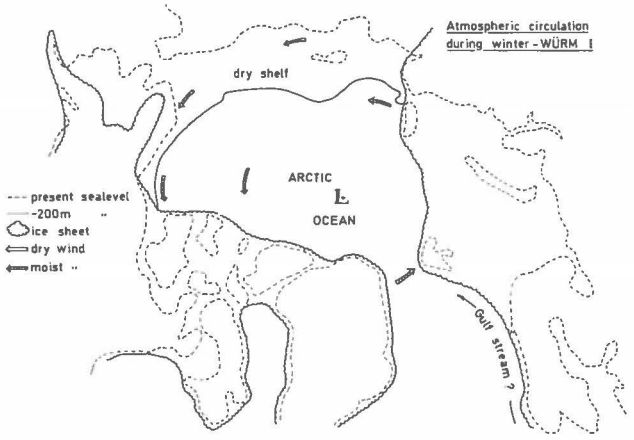
pean rivers, as well as Ob and Yenisey (?) were covered by ice. The water from these rivers, today contributing two thirds of the river supply, was negligible during the glacial advance phases. The other Siberian rivers, likewise draining partly glaciated areas, must also have had a decreased water supply to the Arctic Ocean. On the other hand, the area of the Arctic Ocean decreased by 10–20% due to a lower sea-level. Taking all these facts into account it is safe to say that during glacial advance phases the river supply was reduced to less than 25 % of the present amount. Practically this means a more or less **open Arctic Ocean.** We (Olausson & Jonasson 1969; further developed in Olausson 1972 b,1981, 1985) concluded at that time that the Arctic Ocean – Nordic Seas (= Norwegian and Greenland seas) were practically ice-free during the peak of the glaciations because their surface waters were not stratified (Fig. 28). The only problem with our concept was that the papers were printed some 30 years too early. In very recent papers more and more authors favor an open Arctic Ocean during the peak of the last glacial. The $\partial^{18}0$ diagrams of the Fram cores, published by Marcusson et al. (1985, Fig. 3), and others, show that the surface waters in the Arctic Ocean and the Nordic Seas were isotopically identical during the Weichselian maximum. The $\partial^{18}0$ values of the Arctic cores differed, however, from those in Norwegian sea cores during Alleröd (?) times. Later, the $\partial^{18}0_{seawater}$ decreased by 3 per mil in the Arctic Ocean. It became ice-covered as a result of the huge meltwater discharge and the opening of the Bering Strait. Later studies support the theory of an open Arctic Ocean during the peaks of the last ice ages (see also Olausson, 1995).

No one has produced convincing evidence for an ice-cover over the Arctic Ocean during the Weichselian. Such an ice-cover has been taken for granted, on faith.

The driving mechanism of the surface water circulation of the **Nordic (= Norwegian-Greenland) Seas** is variously ascribed to wind effects, thermohaline processes and the forcing effects of the inflowing sea water entering the sea. The vertical circulation is an effect of evaporation and cooling of the surface layer in winter. Excessive cooling forms dense water masses that overflow the Scotland-Greenland ridge on its way into the deeper part of the North Atlantic basin. The inflow from the north of fresh water into the Greenland Sea is at present 0.08 Sv (1 Sverdrup $=10^6$ m^3 s^{-1}); 0.01 Sv i supplied by melting icebergs in the East Greenland current. The difference between precipitation and evaporation (= P–E) is 0.03 Sv. The total fresh water supply thus is 0.12 Sv (Mosby 1962). Referring to the discussion about the glacial Arctic, we may, for the glacial advance phase, consider the river and ice berg contribution as 0.02 Sv and P–E as 0.03 Sv, respectively, or in total 0.05 Sv at best, i.e., a reduction of the fresh water supply by at least 50 % when compared to the present value. The absence of a salinity stratification in the Nordic seas during the early – peak glacial stages together with the intensified cooling made it impossible to cover these seas by ice during the main part of a glacial. Instead, the Nordic Seas and the then open Arctic Oceans could have been an area of more intensified sink water production , as illustrated in Fig. 25 B.

The arguments for an ice-cover in the Norwegian Sea by CLIMAP and Kellogg have been analyzed and criticized by Olausson (1981). The development of the sea-ice in the pre-Weichselian Arctic Ocean and the last three deglaciations (e.g. the cause of the Younger Dryas cooling) are discussed in Olausson & Jonasson (1969), and in Olausson (1985, 1995). I think it is most likely that the Arctic Ocean was open during glacial ages due to the absence of salinity stratification. No arguments for an ice-covered glacial Norwegian Sea have been found. Berger & Jansen (1994 p. 305) support this conclusion and write: "there is now evidence that the Norwegian Sea was open even during a glacial maximum (at least seasonally), so that advection of warm Atlantic water reached as far as Spitsbergen, where it could provide a source for growth of glaciers". The Younger Dryas cooling was caused by the formation of an ice cover in

the Polar Sea (Olausson and Jonasson 1969, Olausson 1969, 1995), an explanation apparently accepted by many scientists.



29. The atmospheric circulation during the cold phases of the last ice age. The moist winds nourished the northernmost part of the Laurentian ice sheet.

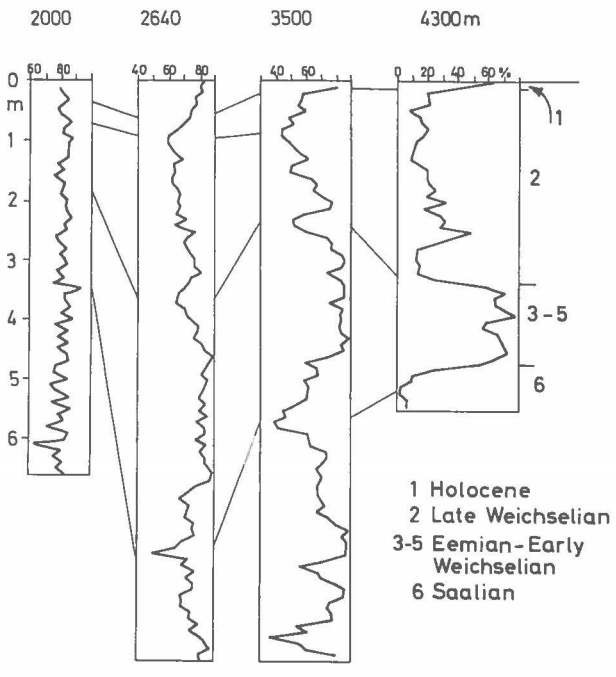
There existed, thus, a large area of homohaline, cold sea water extending from about 40°N into the Arctic Ocean during early – peak glacial times (see also e.g. Haupt et al., 1994, Fig.10b). Lack of stratification and the extremely cold climate on the surrounding continents must have considerably favoured an increased sink water formation (Fig. 25 B).

D. The carbonate accumulation during the late Cenozoic era. The decreased or halted downwelling in the Nordic Seas during the postglacial times (Fig. 25 C–D) resulted in an increased carbonate accumulation in the Atlantic, according to studies of Core 241 A. The highest percentage of $CaCO_3$ in this core was recorded in a layer formed ca. 5,000 B.P. (Wiseman, 1965). Due to lower rate of accumulation in the Indo-Pacific Ocean this trend is not known from there.

The increased downwelling in the northernmost Atlantic during ice ages are seen in from all oceans and especially in the distribution of calcium carbonates.

a) **Saalian-Holocene interval (stages 10-1).** As outlined before, the Weichselian ice age corresponds to the isotopic stages 5 d–2, the Eemian interglacial to 5 e, and the Saalian to the isotopic stages 6–10. The Saalian ice age consists of two main stadials (Drenthe and Warthe) and at least one (Gerdau), two (Rügen and Treene) or more intervening interstadials (see e.g. Flint 1971, p. 628–629; Zubakov & Borzenkova, 1990). These interstadials should appear in a $\partial^{18}O$-curve, at least as strong as the interstadials within the Weichselian stage (5d–3). The Saalian cannot be correlated with a single isotopic valley (6) but only with the isotopic stages 6–10.

It appears from Fig. 30 that the carbonate content in deep sea sediments is constant down to a water depth of ca. 2,000 (–2,500) m in the North Atlantic. The water mass in that interval is the Upper Deep Water which, during ice ages, may get some injection of warm, highly-saline water from the Mediterranean. Below ca 2,500 m and down to about 3,000 m, in the depth stratum of the Middle North Atlantic Deep Water with its origin in the Subarctic Water Mass, the carbonate curves display fluctuations in time. The amplitude of the carbonate fluctuations is twice as large in cores obtained from areas bathed by Lower North Atlantic Deep water (ca 3,000–4,200 m), suggesting an increased production of this water mass during glacial ages. The amplitude of the carbonate content rises to 80 per cent at a water depth of 4,100–4,200 meter or more in the North American and European Basins (Cores 280, 284, 289 and 299). The bottom water at that level is of Arctic origin. We see that largest increase in carbonate solution during ice ages occurred in the sphere of influence of this bottom water. An increased flux of such water could explain the features found.



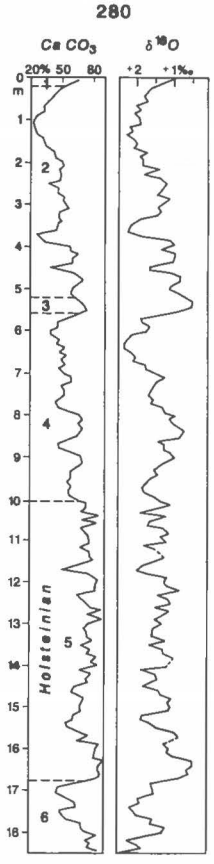
30. The calcium carbonate content of four North Atlantic cores (290, 289, 293, and 299) taken from water depths between 2000 and 4300 m. The isotopic stages 1–6 are given.

When we looked at the cores from the Caribbean we did not observe such fluctuating carbonate curves. That basin is protected against the cold deep waters by a sill reaching up to a water depth of 1,750 m. Even better protected against cold, aggressive bottom water is the Mediterranean by the Strait of Gibraltar, only 320 m deep. When the carbonate in these adjacent seas changes, it has to be ascribed to changing supply of non-carbonate matter.

This pattern with a 0–40 per cent lower carbonate content during glacials than during interglacials is valid for the interval Holocene – Saalian (isotopic stages 1–10). Thus, the circulation during the Weichselian ice age was comparable to that during the Saalian ice age (stages 10–6) in the North Atlantic and also the circulation during Eemian (5 e) and Holocene was much the same.

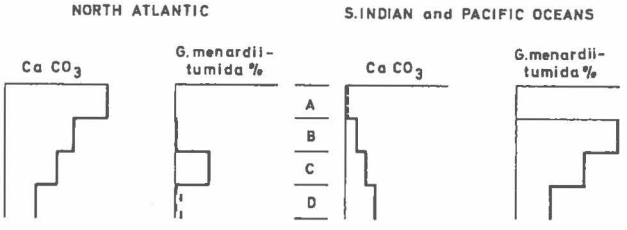
Regarding carbonate content, the stages 1–10 of the Indo-Pacific cores have opposite features as compared with the North Atlantic cores (Fig. 22). During the last two ice ages much more carbonate had accumulated in the Indian and Pacific oceans than during the last two interglacials. The carbonates dissolved in the North Atlantic during the cold stages were deposited on the Indo-Pacific floors (Fig. 25 B). While the CCD was raised some 500 m by the increased supply of "acids" (= H_2CO_3 or $CH_2O + O_2$) in the North Atlantic, it was lowered by about 500 m in the Indo-Pacific Oceans due to this increased carbonate input and rather constant "acid" supply. The bottom water production off Antarctic did not undergo a similar change in magnitude as that in the northernmost Atlantic. The pH of ocean bottom water changed in correspondence with the changed carbonate budget. This pH change has also recently been reported. Analyzes by Sanyal et al. (1995) suggest that the pH in the deep Pacific water during glacial age was 0.3 ± 0.1 higher than at present.

b). **Holsteinian interglacial complex (stages 15-11)** consists of three warm periods with two cool intervals in between. One of the warm periods was about 15,000 years in length (Müller 1974). The cores 234, 235, 280 and ev. 289 (Fig. 30) contain sediments from the Holsteinian interglacial complex, also named the Great Interglacial [= isotopic stages 11–(15)]. Stage 11, here considered as equivalent to the Holsteinian s.s., is characterized by a very high carbonate content in the North Atlantic, higher than in later interglacial stages. The corresponding interglacial stage in the Indo-Pacific cores shows a minimum in the carbonate accumulation. Warm-water indicating foraminifers such as the thick-walled *Globorotalia menardii-tumida* were more or less absent in Cores 234 and 235 (Phleger et al. 1953; Fig. 31), while they have a high frequency in the Indo-Pacific cores. I interpreted the former as a result of dilution by tests of thin-walled species, most easily affected by dissolution (Olausson 1971 a). High abundance of the thick-walled species in the Indo-Pacific cores is due to decreased carbonate sedimentation (Fig. 32). The relative abundance of foraminiferal species in core samples is thus a function of environmental conditions as well as the intensity of the dissolution on the ocean floor. The traditional foraminiferal analysis by e.g. counting "warm" and "cool" foraminifers in the North Atlantic cores have meaning only insofar as the bottom environmental conditions were fairly constant. However, the intensity of dissolution is not constant, as stated above. The dissolution of calcareous matter has not only been more severe than at present in the North Atlantic but also less than



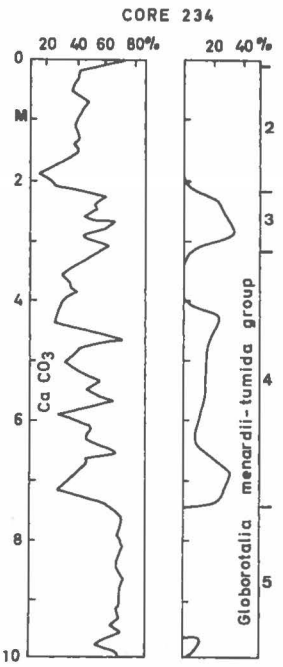
31. The isotopic and carbonate data from the core 280 (34°57´N, 44°16´W; depth 4286 m). Data from Emiliani (1958) and Olausson (1961 d, 1965). The abnormal $\partial^{18}O$-values of the foraminifers between 10 and 16 m were discussed and explained in the text and earlier in Olausson (1988).

at present. During such warm ages a foraminiferal analysis deviates by a still higher percentage of thin-walled species and lower frequencies of thick-walled species, particularly of those produced in small quantities (Fig. 32). This is the trend found in the late Holsteinian age from the central Atlantic cores. In samples from the Indo-Pacific Ocean of the same age the situation is the reverse, supporting the transport theory (Olausson 1965, 1967, 1971a, 1985; Fig. 33). This suggests that the isotopic stage 11 represents a Pleistocene minimum in deep-water production and minimum in dissolution in the North Atlantic, and a minimum in the carbonate production and sedimentation in the Indo-Pacific oceans. Consequently, there was also severe fragmentation of foraminferal tests in the last mentioned stages of the Indo-Pacific cores and a high relative radiolarian content, a fossil group which resists dissolution better than carbonate fossils do.



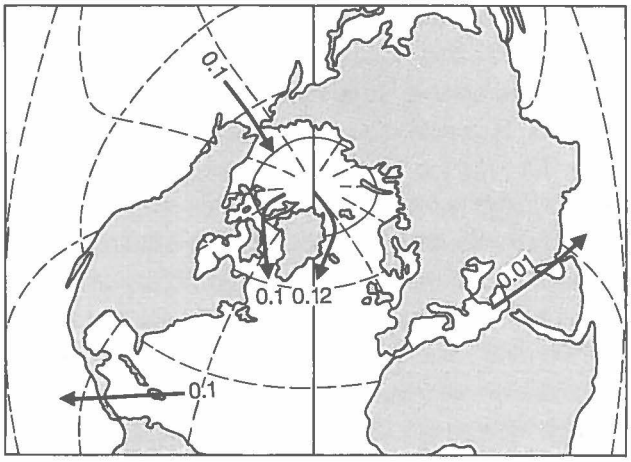
32. Attempt to illustrate the carbonate content and the relative frequency of the foraminifer *Globorotalia menardii-tumida* in North Atlantic and Indo-Pacific cores during
A) if net inflow and outflow of Ca does not occur,
B) during warm ages as (Upper) Holsteinian and periods as Pliocene,
C) during interglacials as Eemian and Holocene, and
D) during glacial maximum.
Note that there is a negative correlation between the rate of accumulation and frequency of the *Globorotalia* group in Indo-Pacific cores. In Central Atlantic cores this group is more or less absent during ice ages due to climatic factors. The group is also absent in stages from very warm ages (B) due to dilution of easier dissolved tests which are produced in much larger quantities than are memebers of the *Globorotalia menardii-tumida* group.

The $\partial^{18}0$ values of foraminifers were higher in the Holsteinian stages than in the Eemian and Holocene sections in our Atlantic cores. They are intermediate between glacial and interglacial values. The (late) Holsteinian sequence was warmer than the later interglacials, with a much lower rate of sink water formation in the northernmost Atlantic Ocean. As stated above, the Albatross core 280 (34°57´N; 44°16´W), analyzed by Emiliani (1958) for the $\partial^{18}0$ of *Globorotalia inflata,* showed abnormal $\partial^{18}0$ values in its Holsteinian stage (see Fig. 31). The $\partial^{18}0$ values are between glacial and interglacial values (Olausson 1965). This must be due to an abnormally high $^{18}0/^{16}0$ ratio in the ambient sea surface water. The water corrections of + 0.95 o/oo to the $\partial^{18}0$ values (Emiliani, op. cit.) valid for the Holocene, are too low for the Holsteinian section, 10–16 m, in this core. The core was taken from an area where the present salinity is about 36.4 o/oo. If the position of the high pressure center were moved northward 5–7° from its present latitude, the water correction, which should be applied to the Holsteinian $\partial^{18}0$ figures, should be 1 o/oo to 2 o/oo higher to make a comparison with the other stages realistic. The $\partial^{18}0$ values of core 280 of the Holsteinian intervals must be near 0 o/oo to be comparable with those from later interglacials. A northward displacement of the zonal belts (by 5–7°) could imply that (a) the Greenland ice cap was reduced in size, at least in southern Greenland, or more or less absent, (b) the atmospheric transport of vapor to the Pacific was far less than at present (see Fig. 34) due to



33. The calcium carbonate content and the frequency of the warm-water indicating *Globorotalia menardi-tumida* group in the 10 m long core 234 from the Sierra Leone Rise (Central Atlantic; from 5°45´N, 21°43´W, 3579 m depth; Olausson, 1960 d, 1965).
The core 0–0.1 m = Holocene); section **2** (0.1–2.1 m) = Weichselian maximum (isotopic stage 2); section **3** (2.1–3 m) = Eemian – Odderade (isotopic stages 3–5); section **4** (3–7.5 m) = Saalian (isotopic stages 6–10); section **5** (7.5–9.68 m) = Holsteinian (isotopic stages 11–12).
Note that the foraminiferal group is absent both in low and in high carbonate stages. Further explanations are given in the text (see also Fig. 32) . – The core below 9.68 m is Miocene in age.

the northerly position of the wind belts, (c) the currents in the Bering Strait may have been reversed. The last in its turn suggests a diminished sea ice extent of the Arctic Ocean due to less fresh water supply to the Polar Sea compared with today, and, thus, warmer condition than at present. Consequently, deep-sea cores provide strong arguments that the (late) Holsteinian subage (stage 11) was warmer than later interglacials, and the export of dissolved matter from the Atlantic was then at a minimum, being in good agreement with the data from other oceans.



34. The fresh water cycle in the northern oceans. The values used are from Weyl (1968) and Olausson and Jonasson (1969). The unit used is Sv (= Sverdrup; 1 Sv = $10^6 m^3 s^{-1}$).

The climatic development during the Elsterian – Saalian interval and the associated hydrographical changes in the Polar Ocean are discussed later on (Chapter XIII).

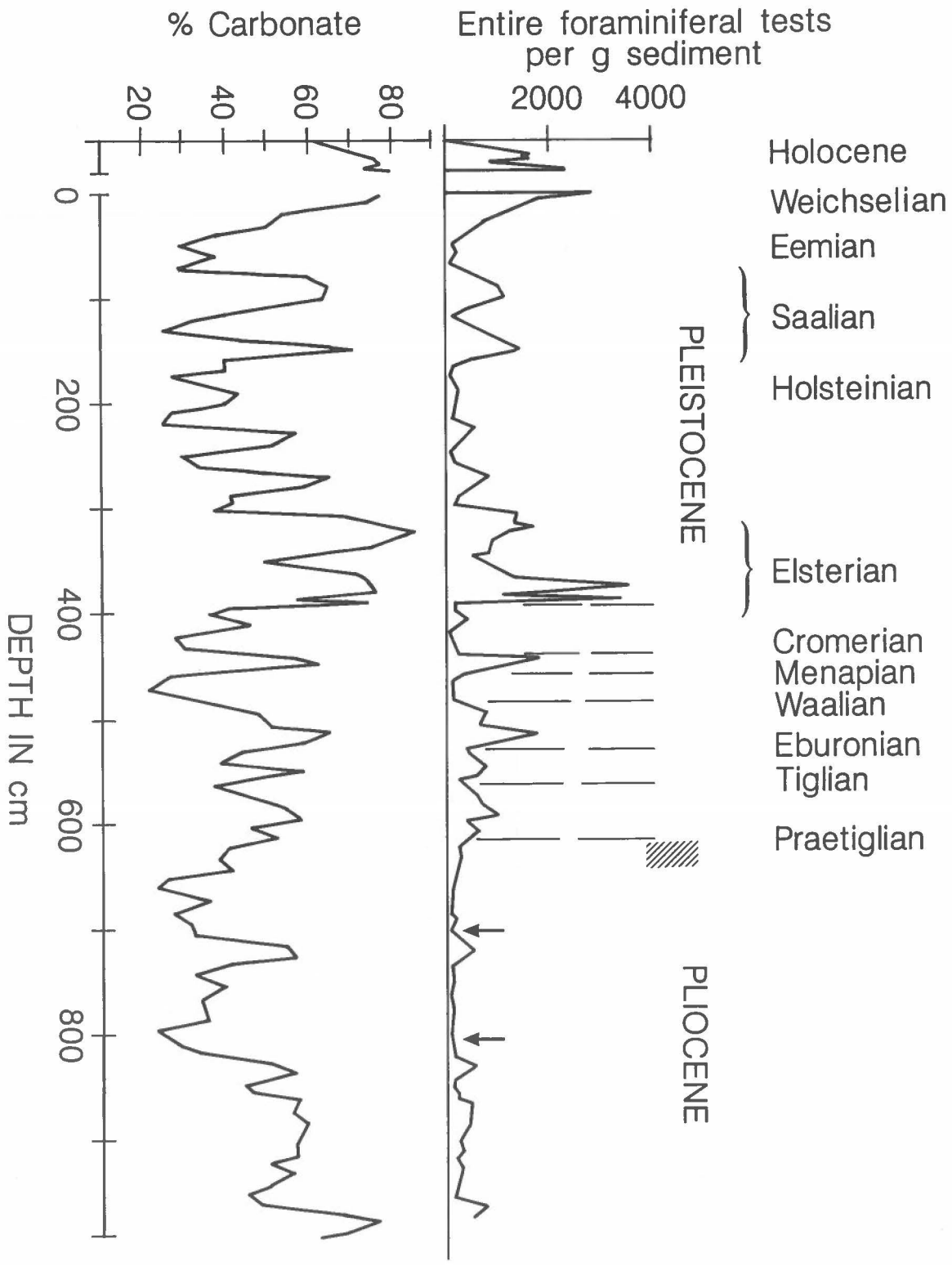
Due to comparatively rapid accumulation in the North Atlantic only the 20 m long Albatross core 280 reaches back to the late Elsterian (isotopic stage 18 or close to the Brunhes/ Matuyama boundary; Fig. 31) while elsewhere, where the rate of accumulation is slower, the cores reach early Pleistocene (Core 154 from the Indian Ocean; Fig. 19–20), and around 2 Ma in the Pacific (cores 58 and 62; Fig. 19, 20, 45, and 46). Therefore, the early Quaternary carbonate deposition in the Atlantic has to be illustrated with DSDP/ ODP cores (see below and Fig. 36).

c) **Early Pleistocene (stages > 15).** The carbonate fraction in deep-sea cores consists chiefly of foraminifers and nannofossils. The former tend to dominate in the more productive parts of the oceans while coccoliths are more common in the oligotrophic areas (see Fig. 21, and Olausson 1961 b). An increase in productivity is reflected by higher numbers of foraminifers in the ocean sediments. Thus, the rate of buried foraminiferal tests per unit of time reflects the past productivity, providing the bottom water circulation has been fairly constant. Such conditions have actually existed in the low-latitude parts of the Indo-Pacific Oceans but not in the Atlantic. Since the rate of accumulation of e.g. entire tests per unit of time is very difficult to determine due to dating problems, the number of tests per gram or cm^3 may be used as an approximation of the burial rate.

It appears from Fig. 35 that the productivity in the Pacific changed through the Pleistocene. This is clear from both the carbonate content and the number of foraminifers per unit of volume. The last is, of course, a rough estimate of the foraminifers/coccolith ratio but it gives nonetheless another information than the $\partial^{18}0$ curves do. Further down in the Albatross core 58 (Fig. 35), below the last occurrence of discoasters (last occurrence for *Discoaster brouweri* is given at 1.89 Ma according to Berger et al., 1993, p. 721), the carbonate content is high while foraminifers are not abundant. This part of the core represents depositional conditions different from those of the Pleistocene, a time of very slow accumulation. A correlation between the Dutch stratigraphy (Zagwijn, 1974) and the deep-sea carbonate stratigraphy as presented is stressed to be tentative. However, the Brunhes/ Matuyama boundary (isotopic stage 19/20) should correspond to the Cromerian/Elsterian transistion, the Jaramillo event to the Eburonian (isotopic stages 27–31; 0.99–1.07 Ma) and the Olduvai (s.s.) is about 1.77–1.95 Ma in age according to the current geomagnetic chronology (Berger et al., 1994). The bottom of Core 58 would thus be about 2.1–2.4 Ma in age.

The Pacific Core 62 extends also back into pre-Olduvai stage and Core 74 back to around the isotopic stages 27–31 (Olausson 1961 b). The longest core from the Indian Ocean Core 154 (Fig.19) has a complete Pleistocene sequence (Olausson 1971 a) from about 2.1 Ma, displaying the same features as Core 58 (Fig. 35). Cores 131 and 152 have sequences up from about isotopic stage 19 (Olausson 1971 a).

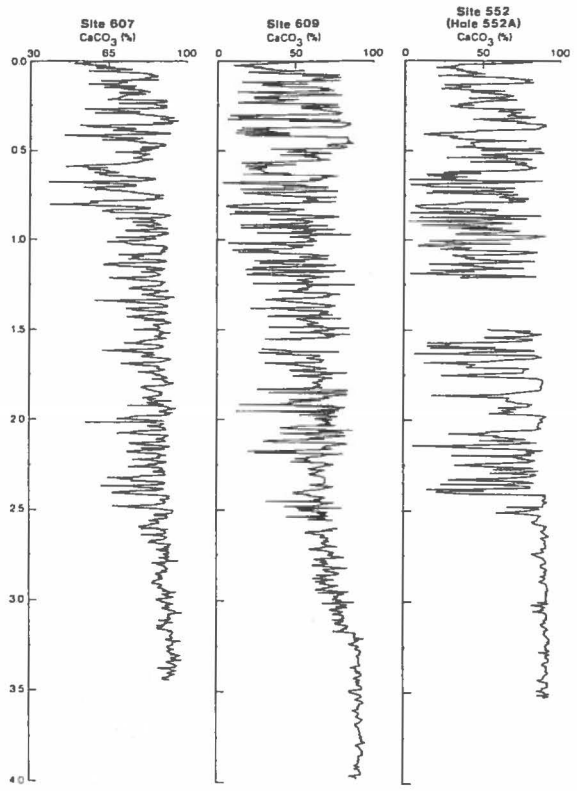
The North Atlantic IPOD-DSDP Sites 552, 607 and 609 (Fig. 36) show a light carbonate content with marked fluctuations back to 2.4 Ma. Carbonate fluctuations beyond this date are less pronounced. It is also obvious that $CaCO_3$ fluctuates less prior to about 0.8 Ma (see especially Site 607) than they do after that date. This date may correspond to the Brunhes/



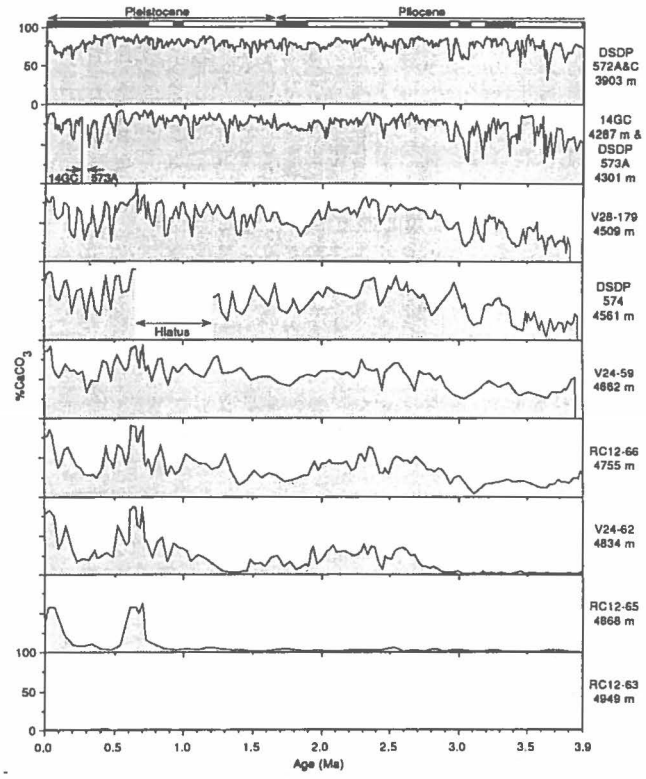
35. The content of calcium carbonate and the numbers of entire foraminiferal tests per g sediment of the Albatross core 58 (from 6°44´N, 129°28´W, 4,440 m depth). Data from Arrhenius (1952) and Olausson (1961 b) with the stratigraphical interpretation by Olausson (op.cit.). The lower arrow indicates the level where most discoasters were extinct, the upper arrow the horizon above which none discoasters are present (Olausson, 1966).

The Brunhes/Matuyama boundary (792 ±10 ky) occurs at the Cromerian/Elsterian transition, the Jaramillo event (0.99–1.07 Ma) during the Eburonian ice age, and the end of the Olduvai event (1.77 Ma) coincides with the lower arrow according to earlier correlation (Olausson and Svenonius, 1975). Quaternary time scale from Berger et al., (1994).

Note that the carbonate maximima in the Pliocene section are caused by coccolithophorids and by foraminifers in the Pleistocene sequence (see Fig. 21). The last is an illustration of the higher productivity in the Pacific due to the increased transport of dissolved matter from the North Atlantic since the last Pliocene (Fig. 25, 36 and 37).



36. Variations in calcium carbonate at the North Atlantic IPOD Sites 607, 609 and 552, plotted againts time (from Ruddiman et al., 1987, Fig. 3). Site 552 is taken from a depth of 2311 m, 607 from 3427 m, and 609 from 3883 m.



37. Variations in % $CaCO_3$ versus age in 10 representative deep-sea cores from the Pacific, arranged in order of increasing water depth from top to bottom (from Farrell and Prell, 1991, Fig. 2).

Matuyama boundary (=Cromerian/Elsterian transition). Between 0.8 and 2.4 Ma the carbonate was generally high in the North Atlantic. This suggests that the carbonate transport to the Pacific was much lower during the late Pliocene – early Pleistocene than later on. Owing to the low import from the Atlantic of biogenous elements during that times the Pacific carbonate sedimentation of late Pliocene age is dominated by coccolithophorids as suggested by Core 58 (Fig. 35). Studies by Farrell and Prell (1991) show that the carbonate deposition in their equatorial Pacific cores, taken from a water depth between 4,700 and 4,900 m are low up to an age of 3.2 Ma. It increases upcore from 3.2 Ma and is rather stable between 2.4 to c. 0.8 Ma. Then the carbonate curves form pronounced peaks during the last three ice ages. Thus, the 2.4 Ma-level (Fig. 37) marks the onset of the present type of the vertical circulation of the oceans. The sinking rate in the North Atlantic was, however, much reduced compared with the present situation until about the Brunhes/Matuyama transition, i.e. to the Elsterian ice age. The reason for these changes in the deep water and carbonate transport will be discussed in Chapter XIII.

A few of the Albatross cores contained also **Tertiary carbonate ooze.** Core 234, from the Sierra Leone Rise in the Atlantic, had below 968 cm a foraminiferal fauna, containing species of upper Miocene age, and below 1112 cm the whole core material was more solidified than upcore, probably formed during the Miocene. The lower part of Core 235 from the adjacent slope was in its lower part of Tertiary age. The core 144 from the Indian Ocean contained also Tertiary sediments in its basal part.

In the central Pacific some cores were raised where Tertiary deposits outcrop or where only thin layers of Pleistocene age rest upon much older sediments. Thus, Cores 69–73, 77–81, 85, 88–91 all have a thin layer of red clay and in case of Cores 90–91 radiolarian resp. carbonate oozes as a top unit, resting disconformably on middle or late Tertiary carbonates (Olausson 1961 b and 1961 c). The transition zone in Core 90 has a thickness of some 64 cm, in Core 76 about 1 m. In Core 69 a sharp (erosional?) boundary is present. In the report (Olausson 1961 c) I discussed particularly the rather fast rise of the CCD seen in Core 90 and argued that other explanations than temperature change or a sudden intense increase of the bottom water production was necessary to explain the feature observed. I suggested that a sudden

local supply of magmatic volatiles set free by submarine volcanic activity was for the time being the most probable explanation. The seafloor around the coring site of Core 90 was thought to have been depressed below the CCD, presumably due to the growth of submarine volcanic cones or shields nearby; two sea mounts are situated not far from the coring site, near the Gilbert Islands. Such a subsidence of the sea floor at Site 90 may have occurred in the lower Miocene during which the sediments of the transitional zone were deposited.

This interpretation of Core 90 was written about 10 years before the introduction of the seafloor spreading theory and the Deep Sea Drilling Project (=DSDP) and 20 years before the discovery of the large oceanic hydrotermal flow. Today we know that east of the core site 90 a spreading axis (East Pacific Rise) from which the bottom moves away by some 10 cm per year with a gradual subsidence of the new floor by some 40 m per million years during the first 70 millions of years (see e.g. Sclater et al.1985). After this time the floor is supposed to spread mainly horizontally. The aforementioned Cores (69–91) are from an area where the bottom was formed during the Cretaceous and had been gradually subsiding while the east Pacific cores (35–62) are collected from a younger part of the Pacific, where the oldest sediments are Tertiary in age. Furthermore, the whole Pacific plate has been moving. The maximum of accumulation along the north-south equatorial Pacific traverse moved during the Miocene from about 5°N to the equator (van Andel et al., 1975). The gradual subsidence negatively affected carbonate accumulation at a certain position through time. The apparent southward movement of the south equatorial divergence (= northward movement of the Pacific plate) thus probably favoured carbonate production in the region now below the present equator.

The duration of the time span when the change from carbonate facies to siliceous facies occurred in Core 90 was earlier given as some 640,000 years (Olausson 1961 c). Using the age-depth subsidence curve by Sclater et al. (1985) we get an average subsidence of 40 m per million years. It implies that this 640,000 is too short to be the only explanation of the facies change observed in Core 90. A temporal increase in hydrothermal activity in the area is another possible explanation. – In the other cores the marked discontinuities made it impossible to attempt an explanation on what happened at the aforementioned coring places during the Eocene – Miocene.

E. A changing $_pCO_2$ and its cause. The carbonate accumulation area of the ocean is at present ca 165 x 10^6 km^2. During the glacial times this area was reduced by ca. 30% in the Atlantic due to an increased carbonate dissolution. However, the carbonate area in the Indo-Pacific Oceans increased by 50% due to an increased supply of biogenic elements resulting in a lowering of the calcium carbonate compensation depth, CCD. In total, the carbonate accumulation area in the World Ocean increased by ca. 20%. Due to the intensified upwellings, discussed in Chapter VI:3 B, the burial of carbon as organic matter increased, as did the accumulation of siliceous organisms.

Compared with recent conditions, the percentages of carbon and $CaCO_3$ are 50–150% higher in glacial sediments from the Indo-Pacific Oceans (e.g. Arrhenius, 1952, e.g. Core 38; Olausson 1960 b). However, a glacial period is a long time span with different intensity, from severe periods to interstadials. A $\partial^{18}0$-curve from deep-sea cores (or from ice sheets as from Greenland) is by far too simple graph of the past temperatures or sea-levels stand, or of the amount of water stored in ice sheets. The oxygen isotope curve of a deep-sea core has normally a peak-to valley amplitude of 1–1.2 per mil chiefly depending on the glacial storage effect (Olausson 1965), and further reflecting the gradual isotopic changes in the Antarctic and Greenland ice sheet (around 0.3 per mil during a glacial; Olausson 1981 b, Mix and Ruddiman 1984). During deglacial times meltwater injections could locally have created large peaks (spikes), which is discussed here in connection with the isotopic data from the Northern Seas, and the Mediterranean (up to 8 per mil deviation as in Moltemyr, SW Sweden; Olausson 1982, see also Olausson 1995, Fig. 1). The last two reasons are responsible for the saw-toothed trends of the North Atlantic ∂18-curves, discussed further below. The extent of glaciation and sea-level change is, thus, difficult to establish from only an oxygen-isotope curve. The real temperature effect on the curve is likewise difficult to determine. Movements of atmospheric high and low pressure centers will also alter the ∂18 of planktic foraminifera as discussed in connection with Core 280. For longer times the hydrothermal flow is supposed to alter the ∂18 of the ancient sea water (Olausson 1988, p.114–117) Here we should recall that the sea water has a passage time through the crust of some 9 x 10^6 years. Isotopically abnormal precipitation with origin in moisture evaporated from meltwater has caused apparent temperature increases

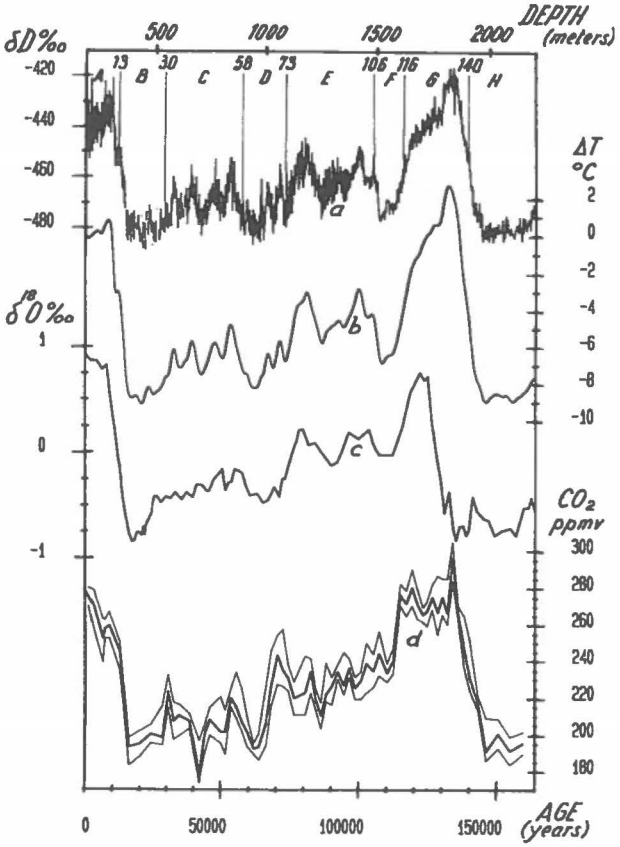
in Greenland ice cores by up to 12°C, while the real temperature change is still unknown (isotopic data of Johnsen et al., 1992). Therefore ice core data must be interpreted critically in the same way as for ∂18 in planktic foraminifers.

A calculation of the excess carbonate deposition for e.g. the whole Weichselian (isotopic stages 5d-2) is not possible to carry out due to aforementioned reasons. Other difficulties are caused by various dating problems both concerning the circulation variability and associated carbonate accumulation during all these subages of ca. 100,000 years (see e.g. Fig. 30). Nonetheless an attempt will be given. Before considering the calculation we must remember, however, that the numbers of mol of carbon dioxide in the present sea is given as 3×10^{18}, and those in the atmosphere is 5×10^{16}. Those in the soil are c. 10^{17} and in the biosphere 5×10^{16} mol, respectively. The total amount of CO_2 delivered to the sea per year is 2×10^{13} mol and the total amount of Ca^{2+} yearly discharged into the sea water is 12×10^{12} mol.

Let us simplify the calculation problem by assuming a doubled calcium carbonate accumulation occurring during 10^4 yrs. (e.g. 25,000–15,000 B.P.). We get $10^4 \times 12 \times 10^{12}$ mol or 1.2×10^{17} mol CO_2 left the oceans in the form of $CaCO_3$ in excess of the normal during this 10^4 years. Furthermore, the accumulation of carbon in other forms should have increased by ca. 20% (= 0.3×10^{17} mol) a figure based on our $\partial\,^{13}C$ analyses of foraminifera. These analyses show a decrease of $\partial^{13}C$ in glacial samples by 0.5–1 o/oo. Since the $\partial^{13}C$ of marine organic matter is ca. –23 o/oo while $\partial^{13}C$ of marine carbonates is ca. 1 o/oo the aforementioned ∂-changes are predominantly (ca 80%) caused by carbonates. Combining the two figures of excess carbon accumulation we find that **ca. 1.5×10^{17} mol CO_2 left the oceans, a reduction of ca. 5% of the total CO_2 in the oceans and a drop in the atmospheric CO_2 pressure by some 50 ppm (cf. Fig. 25 D) during such a long severe cooling period**.

This calculation of the past atmospheric pressure of carbon dioxide was given at the first Hans Pettersson Symposium 1984 (Olausson, 1985, p. 296; another was given by Berger et al., 1988). However, it could had been given already during the late 60´s when the nutrient transport theory was developed even if the C-14 datings of deep sea carbonates at that time were comparatively few. A decade later the large drop in $_pCO_2$ was found in Antarctic ice cores (Delmas et al., 1980; Neftel et al., 1982; Lorius et al., 1988). Later studies of the air bubbles trapped in the Antarctic ice sheet have extended the record to the last phase of the Saalian ice age, or back to about 240,000 years B.P. (Jouzel et al., 1993).

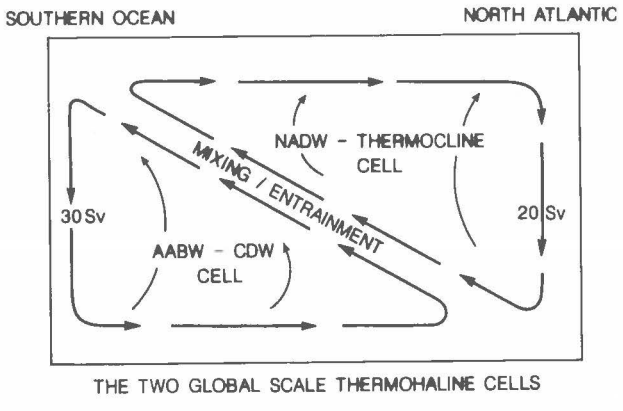
Judging by Fig. 38 (and Fig. 5 in Jouzel et al.´s paper) the CO_2 content decreased gradually during the last phase of the Saalian glaciation (isotopic stage 6), to a minimum of around 200 ppmv. During the following deglaciation and earlest Eemian the carbon pressure increased rapidly to about 300 ppmv. The carbon level remained at 270–300 during the warm Eemian through the temperate Odderade (stages 5e–5a). Then, CO_2 decreased with the same intensity as ∂18 in deep-sea cores increased, down to the same CO_2 as at the end of the penultimate glaciation. From about 16 ka to about 10 ka, which is the deglacial period when the sink water production in the North Atlantic was reduced or ceased, there was a rise in the $_pCO_2$ from about 180 to 280 ppmv. It seems resonable to suppose that this saw-toothed trend of the carbon pressure curve has been a repeated pattern at least dur-



38. Data from Vostoc ice core (Lorius et al., 1988, Fig. 1). Curve a) gives deuterium content per mille versus SMOW, curve b) smoothed Vostoc isotope temperature values, curve c) marine $\partial^{18}0$, and curve d) CO_2 concentrations (ppmv) where the heavy line represents the best estimates. The upper scale gives the depth of the Vostoc ice core, the lower indicates the time in years of the various records.

ing the last ca. 800,000 years, i.e. throughout the Brunhes polarity epoch (the isotopic stages 20-1). The carbon pressure curve had probably a smaller amplitude down to 2.4 or even 3.2 Ma, when the carbonate curves started to fluctuate. If, as suggested above, the variation of the carbon dioxide pressure is caused by changes in the intensity of the NADW-Thermocline Cell (Fig. 39, redrawn from Gordon 1986) with a contemporary rather constant flux of the AABW-CDW Cell (NADW= North Atlantic Deep Water, AABW = Antarctic Bottom Water, CDW = Central Deep Water). The flux of the northern cell increases during glacial times due to cooling of the homohaline water mass in the northern North Atlantic; it decreases to nearly nil when the meltwaters from the adjacent disintegrating ice sheets discharge into the same area at the same time as the surface water is warmed up by the postglacial heat increase. It thus seems that the various CO_2 levels in the atmosphere are a function of the climate by itself and are not the cause for global temperature changes. This is, of course, an important remark during a time when so many speak about an unprecedented warming of the globe. There is no doubt as to the role CO_2 as a greenhouse gas, but the question instead is: **what is acting as a giant global thermostat, helping to stabilize the climate? Is the frequency of clouds the regulator?**

The clouds affect the surface temperature by their effect on both incoming solar and outgoing terrestrial radiation. Daytime maximum temperature are much lower on cloudy days than on fair ones, and cloudless nights are cooler than overcast ones, other

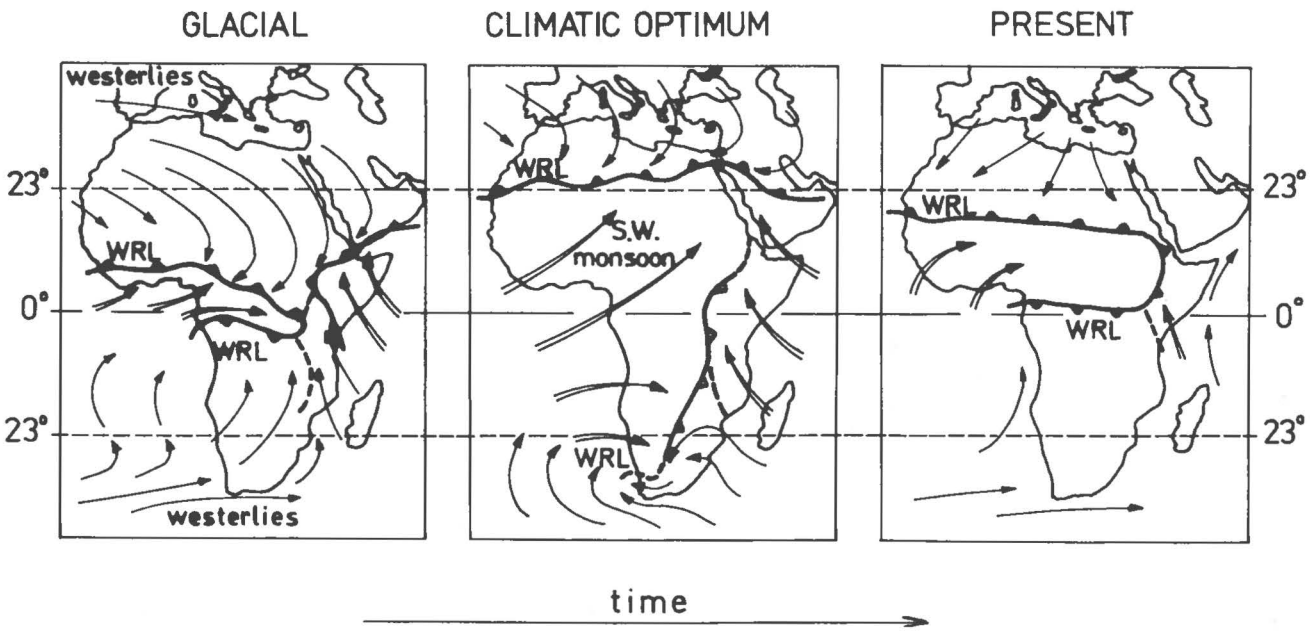


39. Schematic illustration of the vertical circulation of the oceans (after Gordon, 1986). 1Sv = Sverdrup = $10^6 m^3 s^{-1}$. NADW = North Atlantic Deep Water, AABW = Antarctic Deep Water and CDW = Circumpolar Deep Water.

factors being equal. The net effect of clouds at the top of the atmosphere is a cooling of 13 Wm^{-2} while an instantaneous doubling of the carbon dioxide concentration is only around 2.5 Wm^{-2} (Slingo 1990). The average albedo of our planet is about 30 per cent, so only 70 per cent of the incident radiation is actually available to power the climatic system. Clouds cover nearly half the surface of the Earth but account for about two–thirds of the planetary albedo (Crane and Barry 1984). It is beyond my competence to dig further into the cloud´s problems. However, let me only make a few remarks about the theoretical remarks on the presumed thermostat. First, there is general agreement that the temperature was lower during ice age(s), particularly at the glacial maximum, say on average 4°C for the whole Earth. Second, the evaporation changes by 5–8 % per degree (C) which means a lowered moisture (and cloudiness) in the ice-age atmosphere than today (and more during the climatic optimum, i.e. the pluvials). Furthermore, the sky of the cool and temperate mid-and-high latitudes of the globe, is generally covered with more than 50 % (up to 90 %) clouds. Such a high cloudiness is also found in tropical regions (rain forest areas). The dry and semi-dry environments, i.e. the desert, steppes and savannas in between aforementioned cloudy region have a cloud cover less than 50 % (according to WMO). As appears from Fig. 40 the areas covered by e.g. rain forests were much reduced during the peak of the ice ages and the dry areas expanded considerably. Thus, the average global cloud cover, presently about 54 % (49 % for the continents and 58 % for the oceans), might have been reduced during ice ages due to the decreased evaporation, precipitation and a few per cent larger continents owing to the sea-level lowerings. These changes might be the reason why the carbon dioxide-level in the atmosphere can change without marked effect on the global temperature. **The clouds may, thus, be the giant thermostat which keep the global temperature in the range determined mainly by the incoming solar radiation and modified by the Earth´s albedo.**

4. *Foraminiferal studies*

In the introduction to this chapter I mentioned the pioneering work by Fred Phleger and his group (Frances Parker and Jean Peirson) at Scripps. Other important micropaleontologists who worked with the Albatross cores were Wolfgang Schott (Hannover),



40. Three sketch maps of Southern Europe and Africa, to illustrate hypothetical distribution of subtropical fronts and mean July airstreams during a glacial, a climatic optimum and at present. WRL = the Westerly/Trade Wind transition (from Olausson 1971 b).

Ruth Todd (Woods Hole), John D. Wiseman and Cameron Ovey (London), and F. Brotzen (Stockholm). Also the group around Gustaf Arrhenius made important contributions. This applies particularly to Annie Andersson, who also helped me with foraminiferal analyses as long as she was in a good health.

The study of foraminfers in the east Pacific cores was undertaken to investigate the relationship between the concentration of foraminifers and other properties of the sediments with the aim to determine the relationship between production, deposition, and diagenesis. The taxonomical classification of the pelagic foraminifers was not applied in more detail than it could be practised by non-academically trained assistants (Arrhenius, 1952, p. 51). For the purpose mentioned the concentration of foraminifers was recorded by counting the number of specimens in an aliquot part of a 1 g sample. Arrhenius found that the number of foraminiferal tests was doubled at each increase of 10 % of the calcium carbonate concentration, but in extreme chalk oozes (above 80 % $CaCO_3$) the number of tests increased nearly 40 times faster than in the calcareous red clay. Arrhenius thus found that the ratio entire/fragmented tests increased with increasing biogenous accumulation. This was an important observation allowing to study the rate of accumulation and dissolution at the floor of the oceans. Arrhenius made, however, the mistake to illustrate the content of well-preserved (entire) foraminifers using a logarithmic scale. Such a curve caused exaggeration of small changes where the numbers are few, and it underrepresented large shifts when the tests are abundant. I realized this very early on and therefore redrew the foraminiferal data of Core 58 accordingly (Olausson 1961 b). From those on I used normal scales for the data from the other oceans. – For further discussion on dissolution and fragmentation of foraminifers I refer to the previous Chapter VI:3.

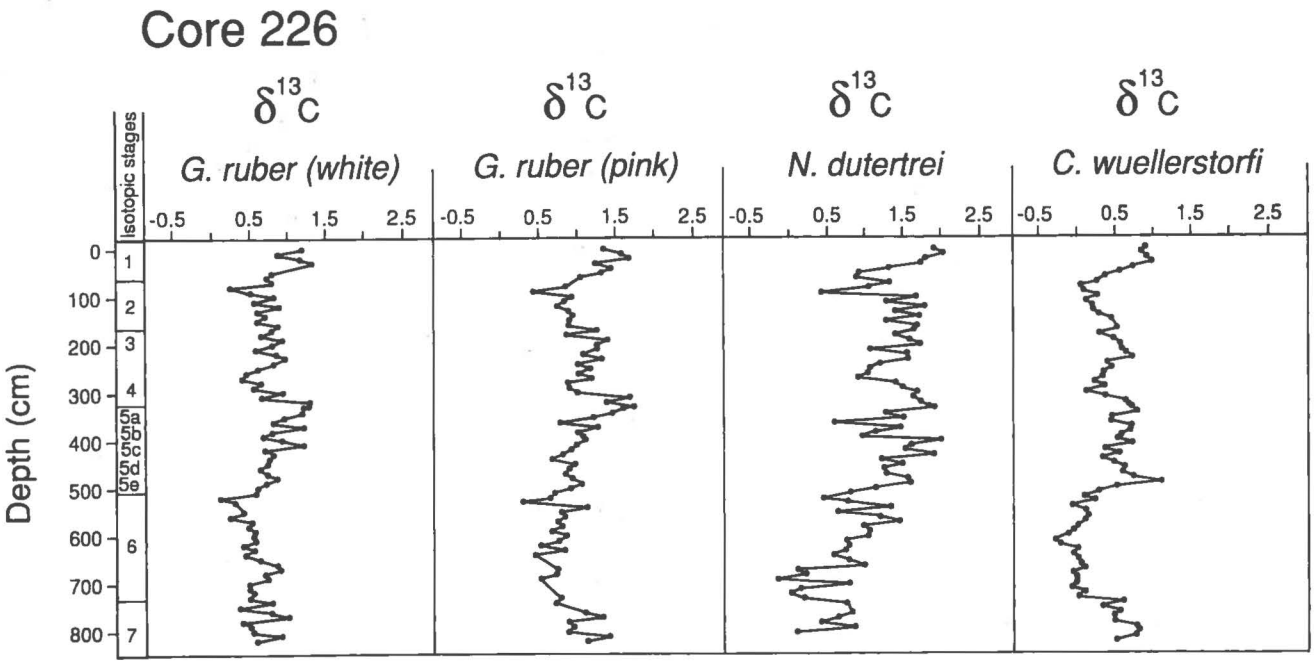
The first worker who systematically used the distribution of foraminifers in pelagic sediment cores for paleoecological interpretation was Schott (1935). He found that the faunal composition recorded in deep-sea cores mirrored well the living assemblages even if dissolution and fragmentation influenced the percentage composition of the thanatocoenosis. A similar result was also achieved by Phleger et al. (1953) in their study of North Atlantic cores. Arrhenius (op.cit.) found an inverse relationship between the number of the tropical genus *Globorotalia* in the planktic fossil assemblages and the total concentration of foraminifers in the sediment samples, a trend earlier noted by Schott (1935) for central Atlantic cores. Furthermore, Arrhenius found that the subtropical species as *Globoquadrina* increased in frequency when the number of tests in the samples increased. These two trends were both explained to reflect changes of the surface water temperature: in-

creased upwelling causes not only lower water temperature but also higher organic production. Three other taxa were also identified, namely *Orbulina universa, Pulleniatina obliquiloculata* and *Sphaeroidinella.* These provided similar paleoecological information as the two large groups mentioned before. Benthic foraminifers generally appeared to be much more resistant than planktic forms, a statement also made by Phleger et al. (op.cit).

Phleger et al. (op. cit.) identified ca. 20 planktic species and divided these foraminifers into five groups according to their environmental relationships. The changes in relative frequencies between these groups were plotted and climatic curves for 39 long deep-sea cores (selected from the North Atlantic Cores 215–296) were reconstructed (op. cit. Fig. 24–26). These valuable data were used by me and complemented with my own data from three other cores and combined with the graphical presentation of the lithology, physics, chemistry and fossil data of the Atlantic cores (Olausson 1960 d, 42 plates). In that report I used *Globorotalia menardii, G. tumida* and *Pulleniatina* as one, warm-water indicating group. *Globigerina eggeri (alias Globoquadrina dutertrei)* and *Globigerinoides sacculifers* represented mid latitude species, whereas *Globigerina bulloides, G. inflata* and *G. pachyderma* were considered as mid-to-high latitude species.

W. Schott (1966) published foraminiferal studies from the cores 22–33, collected 1947 during the initial part of the Albatross cruise across the central Atlantic Ocean, and the results from the cores 215–227 from the second part of the cruise in the Atlantic, off northwestern Africa. Three cores, 22, 29, and 227 were analyzed throughout to stratigraphical standard profiles by determining the planktic foraminiferal fauna in sediment intervals of 0.10 m. Investigation on the benthic foraminiferal fauna gave indication of the presence of downslope transported materials ("turbidites" as in Core 227) as well as information on the frequency of the eolian transport from the Sahara. Core 227 gave a complete biostratigraphical and climatological curve from isotopic stage 11 /10 (earliest Saalian) into the Holocene. Later on a complete foraminiferal and isotopic analysis of Core 228 (Fig. 41–42) was made (Bornmalm, 1988).

The foraminiferal studies of the Indian Ocean cores, carried out by A. Kaipiainen and A. Andersson, were published in Olausson 1960 b, 1971 a, and Olausson et al. 1971. In the first paper I only com-

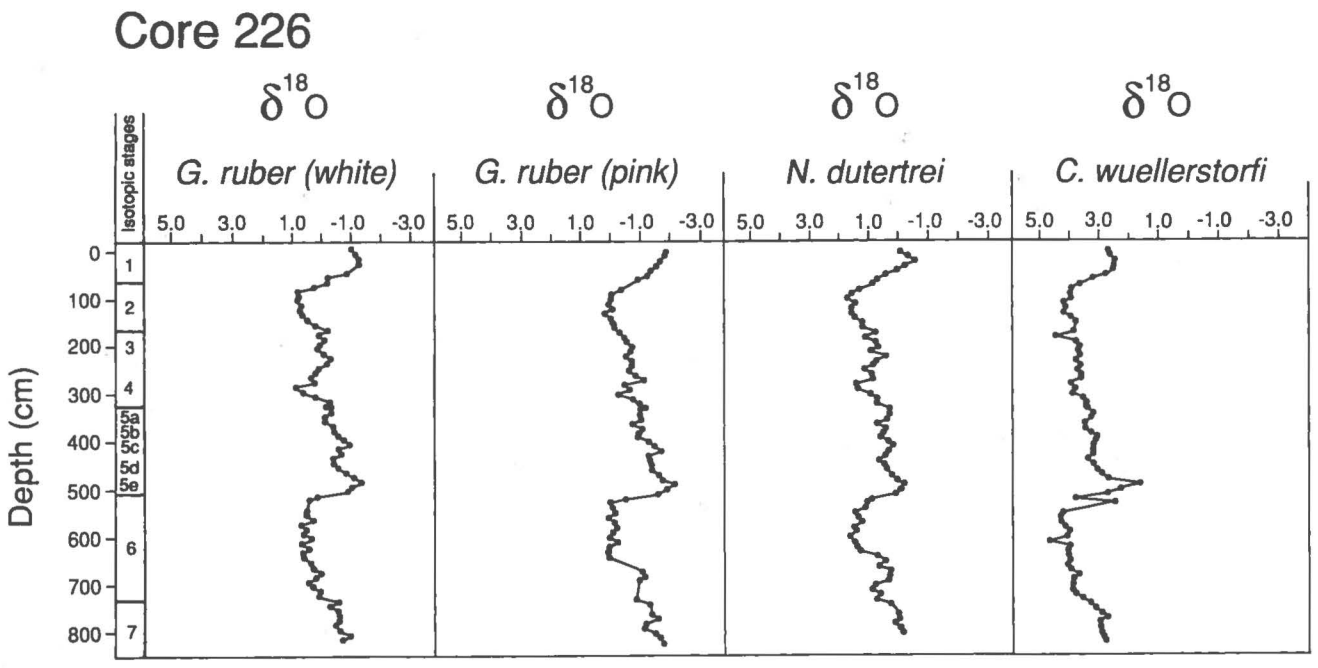


41. Oxygen isotope composition of the planktonic foraminiferal species *Globigerinoides ruber* white variety and pink variety, and *Neogloboquadrina dutertrei*, and the benthic species *Cibicides wuellerstorfi*, expressed as deviation in per mil from PDB standard in the Albatross core 226 from the Cape Verde Island (17° 57´N; 21° 03´W; depth 3100 m). From Bornmalm, (1988).
Note the double peaks at the stage boundary 6/5 for the benthic species suggesting meltwater influence in the bottom water during that long deglaciation, further dealt with in chapter XIII (see e.g. Fig. 86). Note also the $\partial^{18}0$ in the sequence from the Saalian interstadial (stage 7) which shows nearly interglacial values.

Change place with fig 42

Note.
The text to Fig. 41 and 42 at pages 47–48 should change places.

Change place with fig 41



42. Carbon isotope compositions of the planktonic foraminiferal species *Globigerinoides ruber* white variety and pink variety, and *Neogloboquadrina dutertrei*, and the benthic species *Cibicides wuellerstorfi*, expressed as devition in per mil from the PDB standard in core 226. From Bornmalm, (1988).

bined the same foraminifers as did for the North Atlantic cores and I presented the frequencies of the two groups as in the graphs of the plates of Cores 131 – 158. Details of each species and stratigraphical interpretations were given in the two other aforementioned later papers.

The Mediterranean foraminifers were first studied by Kullenberg (1952) and then extensively by Ruth Todd (1958) for the western basin and by Frances Parker (1958) for the eastern basin. Kullenberg studied the cores 189 and 190 for the sediment salinity and the frequencies of the planktic foraminifers, particularly *Globigerina (dubia) eggeri, G. pachyderma,* and *Globigerinoides rubra (ruber).* I will return to his discussions in Chapter XI.

Todd (1958) analyzed 457 samples from 18 cores and found in those only 15 planktonic species in significant abundances. She grouped these foraminifers into warm water, temperate, and cold water indicators and gave the results both in tables and in graphical form. The cores were found to consist of sediments from the last glaciation and the Holocene. The large supply of coarse materials caused a high rate of accumulation. The accumulation in the Eastern Mediterranean was found to be much slower. The cores from that basin were found to penetrate back to the isotopic stages (13)–11 (the Holsteinian) according to my stratigraphical interpretation. Parker (1958) studied 16 long cores and 60 surface samples. The planktic populations in the cores are usually large and show little solution of calcium carbonate tests in contrast to what Phleger et al. found in North Atlantic cores. Parker showed that the faunas reflected climatic temperature variations. The "warm" fauna consists of *Globigerinella aequilateralis, Globigerinoides rubra s.s., G. sacculifera,* and *Hastigerina pelagica.* The "cold" fauna consists of *Globigerina pachy-derma* and *Globorotalia scitula.* The frequencies of *Globigerina eggeri* were supposed to reflect factors other than temperature. I will return to the last problem in connection with the formation of the sapropelitic layers (Chapter XII).

The foraminifers of our Red Sea cores are discussed by Olausson (1971 a) and they are discussed in Chapter XI where also the result from stable isotope analyzes are given.

The foraminiferal analyses in the years around 1960 had the aim to calculate the relative frequency of each species with further grouping of "warm", "temperate" and "cool" species. In that way climatic temperature curves were constructed. The curves reflected not only the surface water conditions but also the bottom water characterics. In a closed basin such as the Mediterranean, the Caribbbean and the Red Sea

Note.
The text to Fig. 41 and 42 at pages 47-48 should change places.

a ± constant pH at the bottom allowed conventional foraminiferal analyses to yield realistic results. However, as had been pointed out earlier by Schott (1935), Arrhenius (1952) and Phleger et al. (1953) the dissolution of tests at the floor complicated the interpretation of foraminiferal data. Dissolution was obviously selective. Their data as well as my chemical experiments presented and distributed in the Cambridge manuscript 1967 (printed in Olausson, 1965, 1971 b Fig. 29.1–29.2) showed an increased resistance against dissolution in the following order: *Globigerinoides sacculifera* > *Globigerina bulloides* >> *Globorotalia (menardii-tumida)* > *Pulleniatina obliquiloculata.*

Before further discussion let us summarize the present classification and the distribution of species in the present-day oceans. There are about 30 planktic species which are grouped into two families: the *Globigerinidae* (spinose forms) and the *Globorotaliidae* (non-spinose forms). According to the modern view the most important species prefer the following waters (Kennett 1982, p.548):

tropical water (24–30°C): *Globorotalia menardii, G. tumida, Pulleniatina obliquiloculata, Neogloboquadrina dutertrei (= Globigerina eggeri* (toothed form)

both tropical and subtropical (12–24°C) waters: *Globigerinoides ruber, G. sacculifer,*

subtropical water: *Neogloboquadrina dutertrei,* (nontoothed form) (= *Globigerna eggeri*)

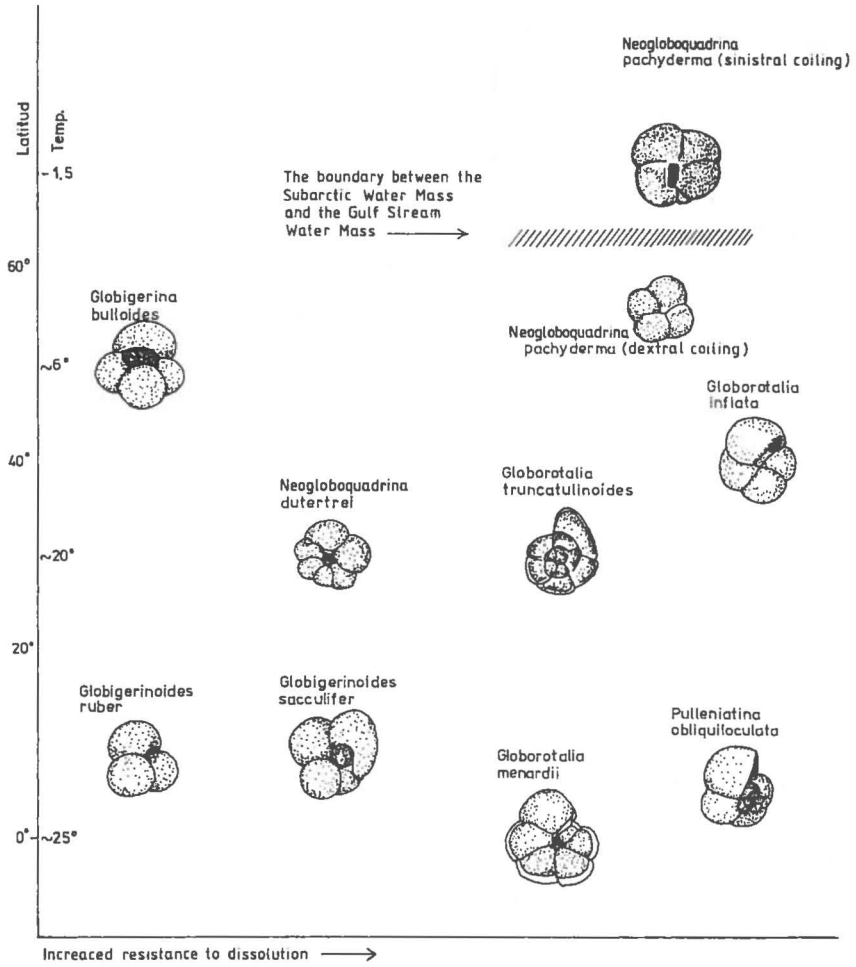
cool subtropical water (12–20°C): *Globorotalia inflata, G. truncatulinoides, Neogloboquadrina (= Globigerina) pachyderma* (dextral coiling)

subpolar water (5–12°C): *Globigerina bulloides, Neogloboquadrina pachyderma* (sinistral coiling),

water (0–5°C): *Neogloboquadrina (= Globigerina) pachyderma* (sinistral coiling).

The vertical distribution of recent (living) foraminifers in the ocean waters is important to know, particularly in stable isotope analyses. Emiliani (1954) and later on Berger (1969) recognized shallow, intermediate, and deep dwellers including the following;

shallow (< 50 m): most spinose species, e.g. *Globigerinoides* and some *Globigerina,*



43. Some planktonic foraminifers distribution in a northern ocean and their resistance to dissolution. Approx. temperatures for the surface waters (and latitudes) are given.

intermediate (50–100 m): spinose species as *Globigerina bulloides, Orbulina universa;* non-spinose species: *Pulleniatina obliquiloculata, Neogloboquadrina dutertrei,*

deep (≈100–300 m): *Globorotalia, Neogloboquadrina pachyderma*

Since the amount of food is largest in the surface waters the rate of production must be highest for shallow dwellers and less for those which occupy deeper habitats.

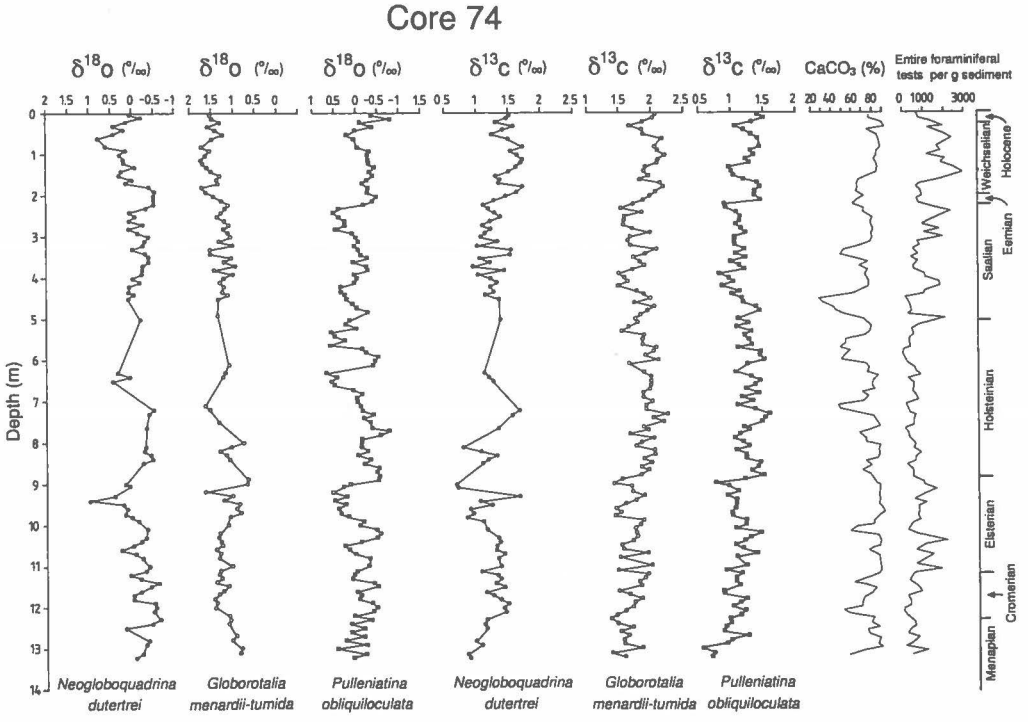
Clearly, that the shallow dwellers can, without any water temperature change increase in relative frequency in cores either due to increased productivity and/or to decreased dissolution intensity at the bottom. Deep dwellers (with thick tests) can, without any water temperature change, increase their relative frequency due to more intense dissolution (more thin-walled species being lost) or decrease in frequency due to increased accumulation of shallow dwellers (higher productivity or reduced dissolution). Examples are illustrated in Fig. 32, 33 and 43.

The classic presentation of foraminiferal data as e.g. Cores 234, 242–246 (Phleger et al. 1953, redrawn in Olausson 1960 d) shows higher frequencies of warm-water species in the Eemian and Holocene sections, much lower in between and downcore. Ericson and Wollin (1956) found similar trends in their six cores from the equatorial Atlantic and the Caribbean but ran into problems in the isotopic stage 11 (pre-Illinoian warm stage or Holsteinian s.s.). This is the same in our Core 234: *Globorotalia menardii-tumida* group was absent in spite of the fact that the isotopic ratio pointed to interglacial temperatures and the carbonate content was high as well (see Fig. 33). I studied this problem and proposed (Olausson, 1965, 1971 b) that the dissolution of carbonate in the North Atlantic could not only be larger than at present, it could also have been less. From the isotopic stages 6–10 to the Holocene the carbonate content had changed from the present value to the glacial ones. The vertical circulation of the north Atlantic was apparently much reduced in the stage 11 (= Holsteinian) compared with the present. During that time the accumulation of thin-walled (spinose) foraminifers must have been increased due to a higher pH at the North Atlantic floor. That was the basic assumption in my discussion about the nutrient transport theory in stages 11–15 (pre-Saalian ages), developed in Chapter VI (3b and 3 d).

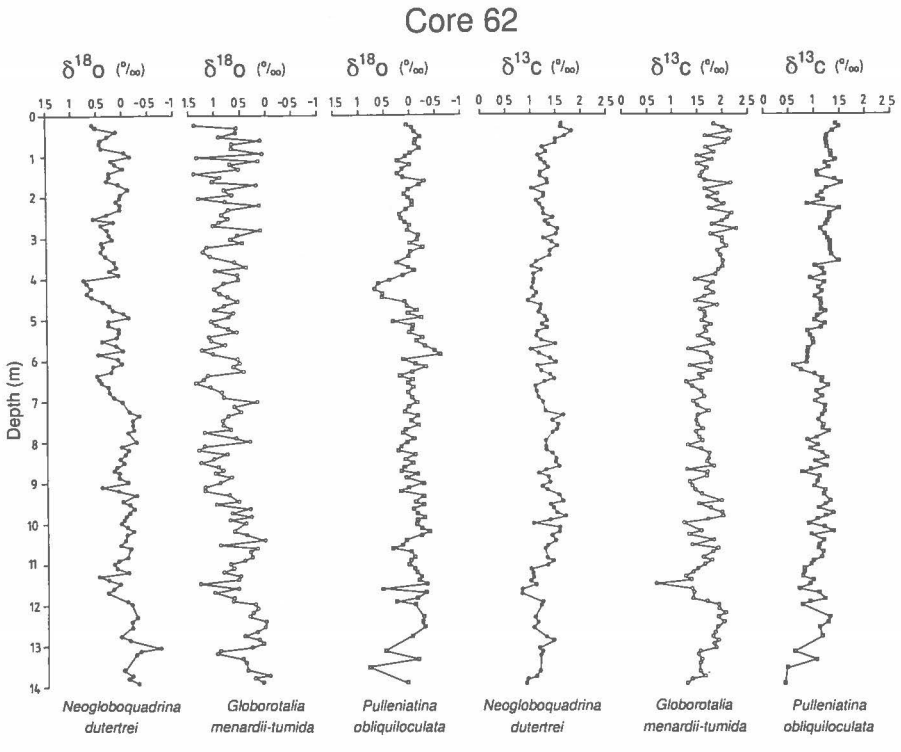
The distribution of *Globorotalia* and *Globigerina* in the Pacific cores showed an inverse relationship (Arrhenius 1952). We have so far no real indication of temperature changes in the equatorial Pacific surface water during the Pleistocene. Since the amplitude of $\partial^{18}0$ oscillations in foraminifers in Cores 58–62 are equal to the amplitude of $\partial^{18}0$ oscillations to be expected in the ambient sea water, the isotopic paleotemperature analysis, supports the concept of a rather constant surface water temperature there (Olausson 1965; Fig. 44–45). Thus the amplitude of deep dwellers such as *Pulleniatina* and *Globorotalia menardii-tumida*) is 1.3 permil in Core 74 down to the stage 11 (the Holsteinian interglacial), below that it is not larger than 1 permil. In Core 62 the amplitude of the same species is about 1 permil through the last 10 stages and about 0.5 o/oo through stages 20–11 (Fig 44–45). The amplitude of the $\partial^{18}0$ of the seawater due to waxing and vaning of ice sheets is calculated to have been around 1.2% upcore from stage 10 and in the stages 16–20 (the Elsterian ice age complex, Olausson 1965), with a smaller amplitude in between and before these times. The changes in frequency of *Globigerina* and *Globorotalia* in our cores can thus be explained as selective solution: more *Globigerina* in glacial stages due to an enhanced productivity, more of the dissolving resistant *Globorotalia* in interglacial stages due to less dilution of the more easily dissolved *Globigerina*.. The bottom water flux into the Indo-Pacific Oceans is, as stated in chapter VI:3, considered to have been rather constant through the Pleistocene. – The foraminiferal studies of Cores 74 and 75 by Brotzen and Dinesen (1959) give the same data as those given by me but our interpretations diverge much (Olausson 1960a, 1961 b).

The foraminiferal analyses from the Indian Ocean can be interpreted in the same way as those from the Pacific. *Globorotalia inflata* reaches a maximum percentage in the stages 16 (–18) in Cores 131 and 128 constituting a good index horizon, used later on also by others, e.g. Jenkins.

The selective dissolution was a problem I tried to overcome by calculating the supposed change in frequency during various degree of dissolving intensity. However, I was not able to find a realistic method to convert core assemblages back to frequencies of the past living assemblages. As stated above the planktic foraminifera are indeed as excellent climatic indicators. One or more species characterize each climatic zone. By counting the relative abundance of the species in the sediment assemblages grouping these to-



44. Isotopic diagram and stratigraphical interpretation of core 74 from the equatorial Pacific (0° 29´S, 151° 33´W; depth 4 545 m).



45. Isotopic diagram of core 62 from the equatorial Pacific (3° 00´S, 136° 26´W; depth 4 510 m). Stratigraphy (Olausson 1961 b, Fig. 4): Weichselian 0–1.1 m, Eemian around 1.25 m, Saalian (stages 6–10) downcore to 2.4 m, Holsteinian (stages 11–15) to 4.1 m, Elsterian (stages 16–20 or 22) to c. 5.8 m. Menapian$_{max}$ (stage 26?) at c. 7 m, Eburonian$_{max}$ at c. 8.4 m, Pretiglian$_{max}$ at c. 10.3 m (the Eburonian ca 1.4–1.6 Ma and the Tiglian ca. 1–6-2.2 Ma according to Zubakov & Borzenkova, 1990, p. 375).
Discoaster brouweri became extinct at 11 m., and 13 m is the upper limit for most discoasters (cfr. the arrows in Fig. 35).
Note the increased amplitudes of the ∂18 values upcore from the Elsterian.

gether in polar, subpolar, subtropical and tropical assemblages, it is thought possisble to reconstruct the original plankton production. The link between living plankton and the fossil assemblage is assumed to be core top samples.

One step further is to calculate the original water surface temperature (and salinity) by means of a **transfer equation**, developed by Imbrie and Kipp (1971). The calibration set consists of core top samples and the corresponding surface water temperatures. The method has been widely applied, e.g. by CLIMAP (see McIntyre et al. 1976). However, the transfer method deviates from previous data processing techniques only by the new statistical procedure. The calibration is based on core top samples where about 50 % of the original tests reaching the floor have been dissolved and, moreover, where the buried tests are partly mixed by bioturbation with older skeletons. These drawbacks must considerably affect the scientific outcome. The variations in abundances of the four foraminiferal groups in downcore samples are thought to give past temperatures and salinities at the position of each coring station. The relation between estimated and measured temperatures and salinities (Imbrie and Kipp 1971, Fig. 21–23) indicate that the errors are large. The CLIMAP map of the Northern Atlantic 17,000 BP (Fig. 27) differs in many ways from what I produced many years earlier (Fig. 31). Their large ice-cover over the Nordic Seas and adjacent North Atlantic probably did not exist (see further Olausson 1981). Besides, the estimated summer temperatures in the stage 11 (Ericson zone U) of Caribbean core V12–122 (l.op.cit. Fig.32) are completely wrong; the method has not been calibrated for dissolution regimes prior to the last two interglacials. This problem was discussed above in connection with the stage 11 of Core 234 and Fig.33 (cfr. also Fig. 32).

The science of paleoceanography, in my opinion, has not benefited much from the introduction of the transfer method. Interpretation of sediment assemblages, according to my experience, is a task more for a geochemically oriented paleoceanographer than a pure micropaleontologist.

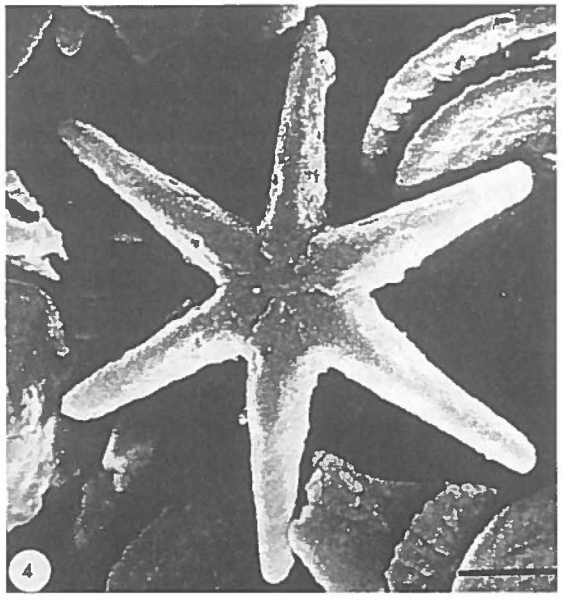
In summary: the studies of foraminifers in sediment cores are useful for dating purposes. A geochemist, trained in oceanography, can use sediment assemblages to extract information about the physics and chemistry of past oceans and adjacent seas. Foraminifers are excellent objects for isotopic analyses. Analyses of the calcium carbonate content, $\partial^{18}0$ of foraminifers, and simple counting of entire tests, the content of weight percentage fraction > 62 μ (or >150μ), as used by Emiliani, are still the most informative methods for reconstructing details of the climatic development.

5. *Notes on*

THE CALCAREOUS NANNOPLANKTON

The coccolithophorids and the related discoasters form a large fraction of recent and fossil carbonate sediments. In some low-productivity areas such as the Mediterranean they constitute the main part of the carbonates, while in more productive water foraminifers can dominate (see Fig. 21). The discoasters are extinct, and the last occurrence of certain species provide a useful stratigraphical datum. The last living discoaster species, *Discoaster brouweri,* became extinct near the Pliocene/Pleistocene boundary, more exactly at the lower boundary of the Olduvai event (isotopic stage ≈ 71 or at approx. 1.95 Ma according to Berger et al., 1994).

I recognized the aforementioned horizon in our Core 58 where both most discoasters are absent and where all *D. brouweri* disappear (Olausson 1966, p.668; 1967, Fig. 2; 1971 a, Fig. 29.9). These two events of extinctions (Fig. 35) happened after two periods of increased carbonate accumulation in the Pacific. A possible link between these changes in the productivity and the extinction of discoaster is not known to me, at present.

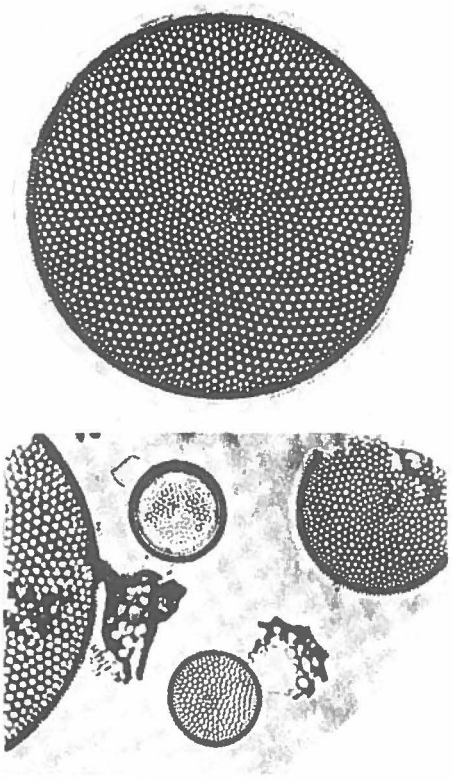


46. *Discoaster brouweri* (reproduced from Stradner, 1973, Plate 34: 4).

6. Evidence from Diatoms and Radiolaria

The studies of diatoms and radiolaria in the Albatross cores were carried out by the two experts of each group: Rudolf W. Kolbe (Stockholm) and William R. Riedel (Adelaide), who after a year in Göteborg, went to La Jolla.

Diatoms. When G. Arrhenius investigated the cores from the equatorial part of the East Pacific he found that a great centric diatom dominated in certain levels of the cores. Arrhenius asked Kolbe to determine the species. Kolbe found that the species was *Coscinodiscus nodulifer* A. Schm. They further found that larger specimens dominated at certain intervals alternating with layers containing only smaller individuals. Arrhenius proposed a biometric survey of the total material with regard to the size variation of *C. nodulifer*. The unique material called for further investigations of the diatom flora and Kolbe was invited by H. Pettersson to make a general survey of the diatoms of the equatorial Pacific cores. This in turn led to a similar survey of the equatorial cores from both the Indian and the Atlantic Oceans.

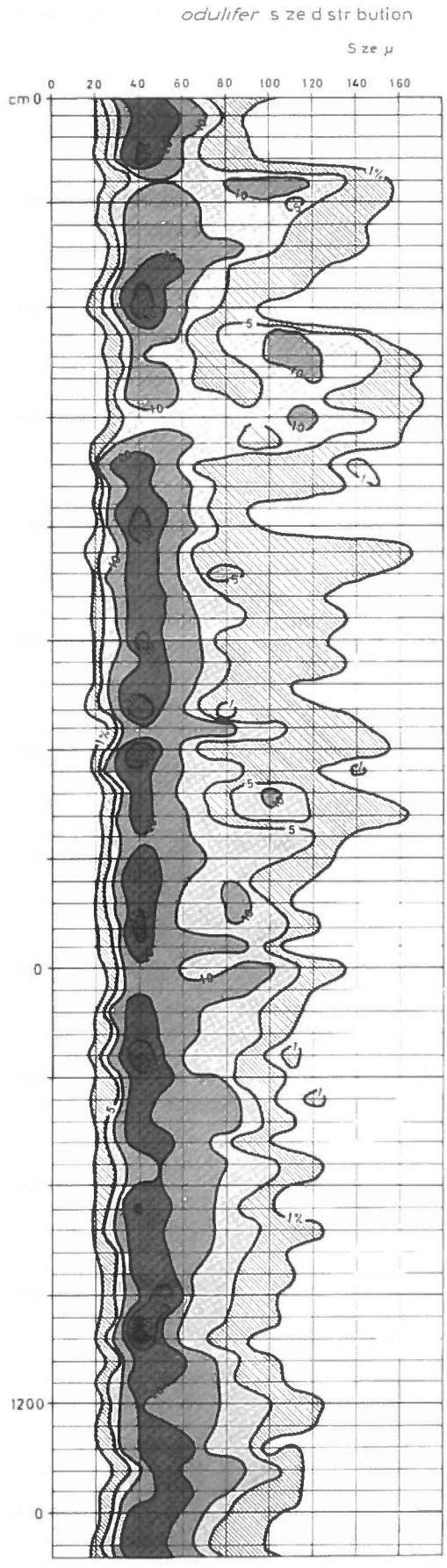


47. Two different sizes of the centric diatom *Coscinodiscus (Azpeitia) nodulifer* A. Schm. in Core 74. From Kolbe, (1954).

The size distribution of *Coscinodiscus nodulifer* became the main subject in the diatom studies. The team who made this study was supervised by Kolbe. Approx. 340 samples from the East Pacific were investigated, corresponding to approx. 120 000 measurements. Their results were summarized in isopletic diagrams of the cores 38, 39, 59, 60, 61 and the upper part of Core 62. These graphs were published by Arrhenius (1952). The diagrams of Core 74 and 76 were placed at my disposal and also published (Olausson 1960 a). All these cores with the diatom data were raised from the sea bed below the Equatorial Counter Current or the adjacent zone to the south. All graphs showed that this centric diatom increased in size in the last two glacial stages and were on average up to 40μ smaller during the intervening interglacials. Arrhenius (op.cit, p. 71) discussed the cause of these size variations.

Diatoms reproduce through cell division. The two valves separate. Both the large valve (epitheca) and the smaller one (hypotheca) secrete new hypotheca. Each generation produces two daughter cells, one of the same diameter as the parental cell, the other of slightly smaller diameter. Thus, in each generation one daughter cell of reduced size is formed. This reduction in size continues until a critical size is reached, when an auxospore is formed. This auxospore forms a new frustule, larger than the parental frustule. If when the minimum is reached an auxospore is not formed, the cells, will die.

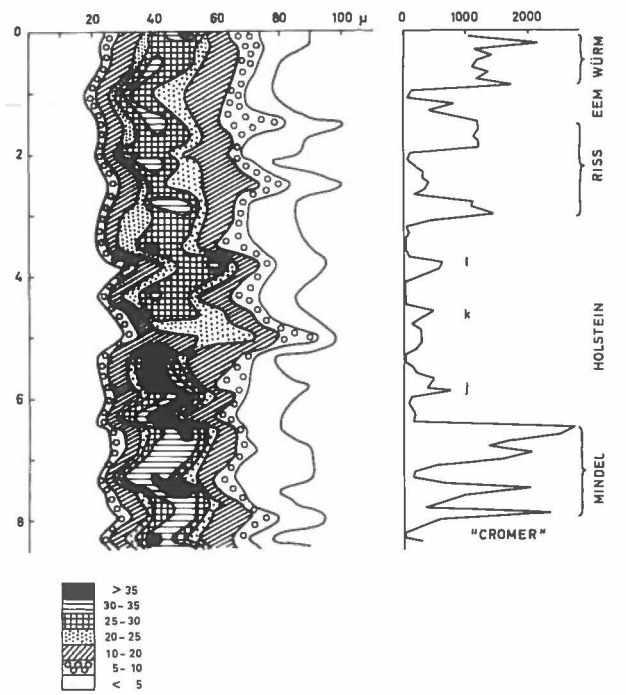
The large-size frequency maxima in the size distribution of *Coscinodiscus nodulifer* were found to be related to periods of high rate of production, not only of diatoms but also of radiolaria and foraminifers (l. op.cit.), particularly during the late Quaternary period. The improvement of life conditions for the species during certain periods may result in an increased production of auxospores (Kolbe 1954, p.11). Arrhenius (op. cit.) found further that the accumulation of this large-size diatom increased by a factor of eight from the stage 5 e to 6. This is entirely due to a corresponding increase in the rate of sedimentation. Similar increases were noted for other centric and pennate diatoms. Arrhenius also suggested that the rate of accumulation of these siliceous frustules gives a still better record of surface productivity than is shown by the rate of $CaCO_3$ accumulation (l. op. cit., p.73). In a few equatorial Atlantic cores Kolbe (1955) found a changing frequency of diatoms only for the stage 10–1 interval. The size of *C. nodulifer* was



48. Size distribution of *Coscinodiscus (Azpeitia) nodulifer* in the Pacific core 74. The stratigraphic interpretation is given in Fig. 50. From Olausson, (1960 a).

smaller than in the East Pacific and only in a few rare cases frustules with 100–125 μ in diameter were observed. Kolbe (1957) studied the Indian Ocean cores as well. *C. nodulifer* was here "never as predominating as in the Pacific sediments" "Periodical fluctuations in size of the whole populations of *C. nodulifer* were never observed in sediments from the Indian Ocean as opposed to the characteristic size variations of *C. nodulifer* assemblages in Pacific cores" (l. op. cit. p. 30). He suggested that the differences between the two oceans as to the size variation of *C. nodulifer* depended on differences in the current systems: the Indian Ocean has no distinct Equatorial Current System as the Pacific has. I came, however, to the conclusion that Kolbe´s observation outlined here given could not be correct.

The auxospore formation in an early stage of the vegetative part of the reproduction cycle is probably favoured by an increased supply of nutrients. These biologically important elements may be phosphorus, nitrogen and/or silicon. The North Atlantic apparently receives 40 per cent of all silica from weathering and from the hydrothermal vents (Indian Ocean 29 %), but a part of this must be lost in the same way as phosphorus, nitrogen and calcium are. It is therefore not necessary to distinguish between N, P, and Si since all will promote the production of siliceous organisms. We may expect to find in the southern hemisphere of



49. Size distribution of *Coscinodiscus (Azpeitia) nodulfer* in the Indian Ocean core 131. Modified after Olausson, (1971 a).

the Indian Ocean and the Pacific the same cycles for acccumulation of organic silica as for carbonates. Although not reported by Kolbe my transport theory suggests that similar cycles in the *Coscinodiscus nodulifer* should be found in cores from the area around the South Equatorial Divergence in the Indian Ocean. I asked a student of mine, Gunvor (Karlsson) Larsson to carry out a similar study of the Indian Ocean core 131 as Kolbe´s group had carried out of 7 central Pacific cores. She found as predicted the same rythmic variations in the size distribution of *C. nodulifer* as earlier had been found in the Pacific cores. The difference between Stages 10-6 and Stages 5 d-2 is not as marked as in the Pacific, but the most pronounced size maximum is again found at my index horizon "l" (= isotopic stage 12; the index explained in Olausson 1961 b, 1971 a). Moreover, the 1 per cent-line reaches highest size values during cooler phases after "k" (=isotopic stage 14), both in the Indian and the Pacific Oceans. This fact also supports the cross-correlation suggested above, based on other data, and is indicative of larger size amplitudes of the diatoms upcore, from "k". Below "k" the size amplitudes are limited, which appears especially in the frequency area of > 35 per cent in Core 131. The diagram for Core 131 shows more marked size fluctuations in the stages 20-16 (= the Elsterian ice age complex) than is found for the Pacific cores. Above "k" the productivity changes are better marked in the Pacific cores than in the Indian Ocean core; below "k" the reverse may possibly be the case.

Scarcity of diatoms in Central Atlantic cores (e.g. Core 235) has made it difficult to carry out such an investigation there.

The size changes in the diatoms were subsequently discussed by others. Belyaeva (1972) showed a relationship between diatom valve diameters and the rate of upwelling, with increased size correlating with increased upwelling.

The changing size of *Coscinodiscus nodulifer* particularly was discussed by Burckle and McLaughlin (1977). The feature of bimodality and the use of this property in biostratigraphy and paleoecology was dealt with. Both in their study and in the data from Arrhenius (op.cit.) bimodality occurs south of the Counter Current (south of 5° N). Burckle et al. (1981) concluded that in the equatorial Pacific the size of *C. nodulifer* was larger in strata from 18 Ka than in State 5 e layers (= Eemian): A fact clearly appearing from the aforementioned isopleths from the Albatross cores, but definitly not visible in their own diagram (Burckle et al., 1981, Fig. 2). Their diagram is difficult to use for a proper conclusion. Arrhenius (1988) also returned to the problem by mentioning the existence of two distinct varieties of *C. nodulifer* which was previously dealt with by Kolbe (1954, p.11). His conclusion was that the auxospore hypothesis is more probable than the aforementioned alternative. Since the size of *C. nodulifer* also changes in the Indian Ocean cores – which apparently neither Kolbe nor Burckle et al., were aware of – I suggest that increased supply of nitrogen, phosphorus and/or silicon to the Indo-Pacific Ocean as the dominating factor favouring the growth of larger species.

Lisitzin (1985, Fig. 11) showed a gradual increased silica deposition in the equatorial Pacific during the Brunhes polarity period. Lange and Berger (1993) have studied the diatom productivity in cores from the Ontong Java Plateau. They likewise found that *Azpeitia nodulifera (=Conscinodiscus nodulifer)* exhibited significant size variations related to glacial and interglacial stages, and, judging by their Fig. 5 A, particularly during last 10–12 isotopic stages (the Holsteinian – Holocene intervals as was observed in the Albatross cores.

According to Defant (1961) the circulation in a fully symmetrical ocean exhibits two closed circulation systems. The boundary between each hemisphere occurs below the Equatorial Counter Current (see also Olausson 1971 a, Fig. 29.4). Thus the increased supply to the ice-age Indo-Pacific oceans would stimulate predominantly the productivity and increase the size of diatoms south of ≈ 4° N. In Chapter XIII I will discuss why the diatoms flucuate in size only during the last ten stages (= post-Holsteinian Pleistocene).

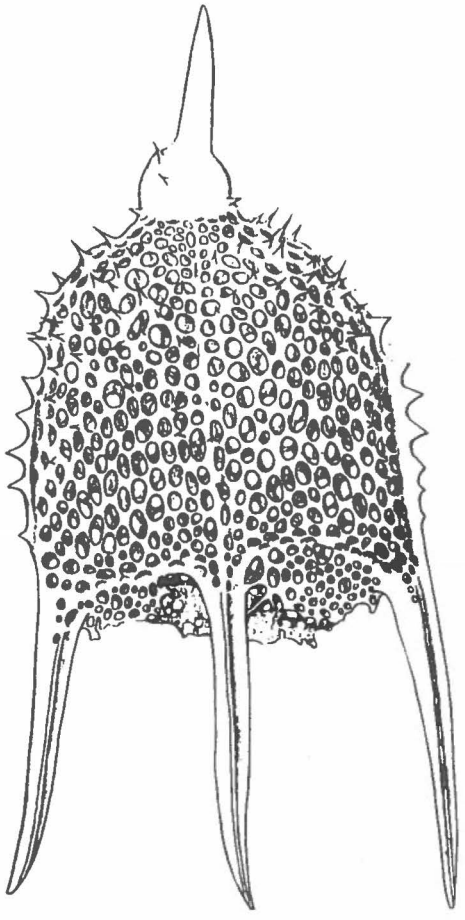
Coscinodiscus (Azpeitia) nodulifer(a) was found in most samples of all investigated Pacific cores. Among the other diatoms *Nitzschia marina* is the second in abundance. Tertiary diatoms occur in Core 58 and particularly in Core 76. Both species are tropical ones. The uppermost 59 cm of Core 76 consists of redeposited matter. Down to about 5.2 m the sediment is of Pliocene age. Below that level, in the Miocene strata, *Coscindiscus nodulifer* and *C. excentricus* are absent (Kolbe 1954, Table 4; Olausson 1961 b).

Kolbe (1957) found what he called ***Ethmodiscus ooze***, sediments consisting "almost exclusively" of fragments of *E. rex*. This "ooze" belongs in my classification to the less calcareous or non-calcareous deep-sea deposits into which fragments of *E. rex* are inter-

spersed. He found such layers in Cores 128, 131, 133, 147, 153 and 234. So far as I understand are the fragments in question redeposited Tertiary diatoms, occurring in Weichselian strata (234), or other Pleistocene layers (131, 153) and in pre-Quaternary sediments (128, 147). Mikkelsen (1977) made a new survey of the origin of *Ethmodiscus* ooze and gave five hypotheses. Redeposition (winnowing) seems likely since the "ooze" occurs when other materials are removed as in the carbonate minimum in Core 234.

Kolbe (1955) found **freshwater diatoms** in a few Atlantic cores, and particularly at 552 cm in Core 234. He identified 17 freshwater species in this layer. The presence of foraminifers (50 % $CaCO_3$) suggests that the diatoms are transported by air or currents from land or estuaries, an interpretation also given by Kolbe. The transport had occurred sometimes during the "Great interstadial" (isotopic stage 7, i.e. in the middle of the Saalian ice age complex).– These freshwater indicators were frequently used by non-geologists (e.g. Nils Hj. Odhner, René Malaise) to speculate about the existence of huge islands in the prehistorical Atlantic Ocean, or whether a part of the midoceanic ridge had been raised above the past sea-level. The evidence was also used to allocate the site where Plato´s Atlantis had existed.

Radiolaria. The radiolarian fossils in the Pacific cores 73–93 were investigated by Riedel (1952, 1957). Three of these 11 cores displayed uninterrupted sequences of Quaternary sediments (74, 75, and 93). In four cores (76, 83, 87 and 91) a rather thin layer of probably Quaternary sediment rested disconformably on middle or late Tertiary deposits. This, and the common occurrence of reworked earlier species in most sediment intervals indicate that a considerable amount of particles had been transported over the floor of the western Pacific during late Neogene. There is no evidence that the cores penetrated into sediments older than middle Tertiary. Five species found were considered as indicators of Eocene, another five as indicators of Oligocene-lower Miocene, and a larger group was considered to contain middle Miocene – Recent forms. Thus, e.g. *Pterocanium prismaticum* (Fig. 50) was a Pliocene species and *P. pratextum* appeared in middle Pliocene and is still living species. – Riedel showed by his Albatross studies that the radiolarian fossils are stratigraphically useful in cores consisting of radiolarian and siliceous oozes.



50. The radiolarian *Pterocanium prismaticum* from Core 73 (211.5–212.5 cm). From Riedel, (1957).

VII. Cosmic dust and the red clay

If we look skywards on a cloudless and moonless night in August we may see streaks of light suddenly crossing the heavenly vault and with equal suddenness become extinguished. These are the so-called "shooting stars", cosmic fragments entering our atmosphere with tremendous speeds. Friction against the rarified air 100 to 200 km up, converts the greatest part of their enormous kinetic energy into heat and light. Most of the particles entering our atmosphere are very small and they become vaporized and dispersed as meteoric dust in the atmosphere. When larger particles, meteorites, pass through our atmosphere the friction raises the temperature of the their surface to incandence. With larger dimensions only the surface layer reaches the temperature of fusion and in case of an iron meteorite it is torn off due to friction in the shape of small droplets of molten nickel-iron. These small droplets consisting of an iron-nickel core with a magnetite coating formed by oxidation in the atmosphere fall to the surface of the Earth and into the oceans. In the latter case they are coming to be embedded into the ocean floor together with terrestrial and marine material. Already the Challenger expedition had recognized these particles and called them "cosmic spherules" (Murray and Renard, 1894).

51. Professor Hans Pettersson with his electromagnetic extractor, and the M/S "Albatross" in the background.

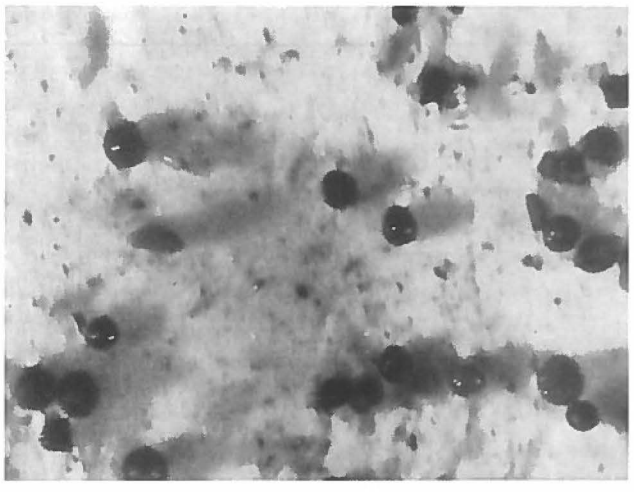
Pettersson (1958) estimated the amount of cosmic dust (micrometeorites) to be of the order of 3.6 million tons/yr. His estimate was based on the amount of magnetic spherules reaching the Earth´s surface. However, this estimate was supposed to be too small since only magnetic particles can be identified as cosmic dust. The vast majority of meteorites has, however, a stony composition of a predominantly non-magnetic character. Later estimation of the accretion rate of extraterrestrial material shows much lower values than those quoted above, i.e. 0.3 million tons per year as presently given.

The accumulation rate of red clay is of the order of only 1 mm per 10^3 yrs, that of the radiolarian ooze ca. ten times larger and the foraminiferal oozes 2–3 cm per 10^3 yrs. Pettersson focused his interest on the magnetic spherules, pellets rarely exceeded 0.2 mm in diameter. He aimed to use the frequency of these grains both to estimate the rate of accumulation of deep-sea sediments, and to see if there had been any changes in "cosmic rain" back in time. Such a change could indicate that the whole solar systems had been passing through interplanetary clouds in the near past. With this in mind he and Kullenberg had developed a piston corer with a larger diameter than the ordinary corer (90 mms in diameter instead of 46 mms). This corer would give him a four times larger volume of sediments per unit length of core. On the other hand the maximum length of such a thick core must be necessarily reduced to 5–6 m instead of the 15–20 m length obtainable with the Kullenberg corer (Pettersson 1955). Unfortunately, the only large-diameter corer was lost in the first attempt and could only be replaced after they had reached Honolulu. Therefore, large-diameter cores were not collected in the eastern Pacific but three were taken in the western Pacific, three in the Indian Ocean and seven in the Atlantic.

While Murray extracted the spherules from clay suspension with an hand-electromagnet, Pettersson developed new equipment for the extraction, a powerful electromagnetic extractor. This "ferrofilter" from USA consisted of 24 magnetic iron sieves, energized by a series of coils, through which a current of 10 amps. was sent (see Fig. 51). Dry red clay powder

from Core 90 was passed through the magnetic extractor. Most spherules were obtained from the fraction 30–60 μ. The resulting number of spherules per kg red clay is illustrated at Fig. 52, which demonstrates marked variations. Pettersson and Fredriksson (1958) and later on e.g. Utech (1963) published the distribution of these spherules in the Albatross cores.

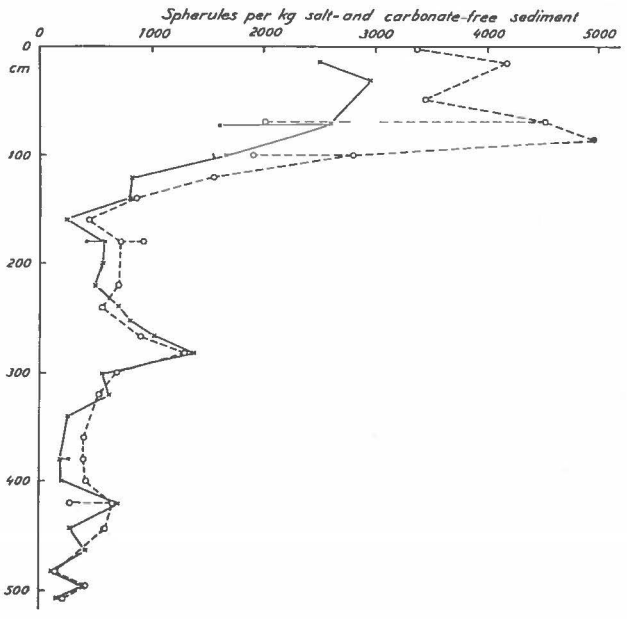
As stated above, a recent estimate of the yearly supply of cosmic dust suggests it to be 0.3 x 10^6 tons. Earlier larger figures had been based on the assumption that all magnetic spherules were of cosmic origin. However, in his attempt to collect magnetic spherules at the summit of Mt. Haleakala on Maui, Hawaii, Pettersson understood that magnetic spherules also were produced by volcanoes (and by weldings!). Counting such particles in areas where products of volcanos and weldings are widespread can give rise to very large errors. If we compare the extraterrestrial contribution with the terrestrial supply to the oceans it can be mentioned that the yearly amount of inorganic matter delivered to the sea is 2 x10^{10} tons and that ca. 3 x10^{12} tons skeletal materials settle down yearly to the ocean bottom. Thus, in the case of red clay the cosmic dust is mixed with a component which is ca 10^5 larger than the dust fraction and in chalk ooze the terrestrial contribution is ca 10^7 larger than the cosmic rain. It was, thus, an impossible task to trace any change in the cosmic dust by studying its distribution in sediments where other components have settled down 10^5–10^7 times faster, and where many of these components are magnetic.



52. Magnetic spherules from the West Pacific core 90. The sperules are about 30 to 60 microns in diameter. From Pettersson, (1955).

It can be added that the discovery of spherules in sediments from the present down to Miocene (Pettersson and Fredriksson, 1958; Fig. 53), suggests that meteor falls have been going on through millions of years and have not been limited to the late Quaternary time. The veils of interplanetary dust would have intercepted solar radiation and thus can reduce the portion the Earth´s insolation, not only during Quaternary but also in Tertiary times.

When this section was written I read the following introduction in Nature (7 Dec. 1995, p. v): "Records of climate spanning the past 900,000 years reveal a mysterious 100,000-year cycle in glacial activity, often ascribed to changes in solar heating due to variations in the Earth´s orbit. But this idea is challenged by the discovery of a similar cycle in the helium-3 record from ocean sediments. As helium-3 is mainly delivered to the sea floor by interplanetary dust particles, the new results suggest that the 100,000-year climate cycle might be linked to periodic changes in the accretion rate of cosmic dust to the Earth." – Hans Pettersson would have been pleased to read this statement.



53. Distribution of spherules in two parallel series of samples from Core 90. From Pettersson and Fredriksson, (1958).
The drop in the concentration is due to a change in the accumulation from chalk ooze to radiolarian ooze core from 258 cm. The chalk ooze is formed ca. three times faster than the radiolarian ooze. The core, Miocene in age, is dealt with in chapter VI:3 D.

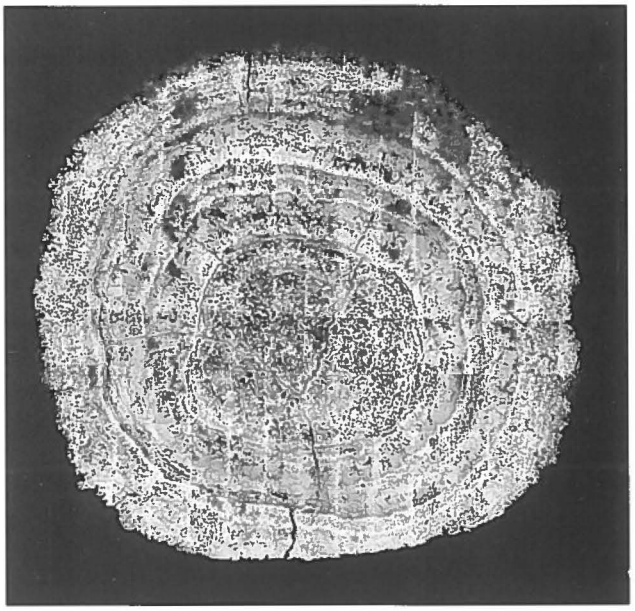
VIII. DATING ATTEMPTS

Pettersson had a great interest in nuclear physics. Between 1922–1936 he worked periodically at the Institute für Radiumforschung in Vienna and spent some times also with Sir Ernest Rutherford at the Cavendish Laboratory, Cambridge. Particularly famous became the papers he wrote together with Gerhard Kirsch in Vienna, e.g. "The artificial disintegration of elements" (Arkiv för matematik etc. 20 A,16, 1927). Also after he had became professor in oceanography in 1930 in Gothenburg, radium was still one of his main interests. Pettersson wanted to explain the relative deficiency of thorium and radium in sea water when compared to uranium.

Pettersson (1937) had proposed the hypothesis of sedimentation of thorium and ionium (Th^{230}) together with the carrier ferrihydroxyde. If this assumption was correct the elements of the uranium family in the uppermost sediment layers should not be in radioactive equilibrium (Kröll, 1955). Piggot and Urry (1941) made use of this hypothesis to date deep sea sediment layers and obtained values for the rate of sedimentation of the red clay and the foraminiferal ooze (l. op. cit.).

The Albatross expedition collected cores in order to pursue further such investigations. Slices were sampled from thirteen cores, mainly red clays, in order to plot curves of the radium distribution. During the course of examination it became evident that the problem of radium in the sediment was more complicated than could be foreseen from Piggot´s and Urry´s investigations (Kröll, op. cit.). Thus it appeared that understanding of the geochemical behaviour of the radioactive elements is the essential basis for a successful interpretation of the results. Kröll´s work did not claim to be conclusive. Also the attempt to date the rate of growth of the nodules was not successful because of the solubility and migration of radium in the nodules (as in the sediments).

On the deep sea floor, where the rate of accumulation is not large, curious products of the chemical processes called "manganese nodules" occur. It is an authigenic concretion built up around a nucleus, e.g. a shark´s tooth or fragment of pumice. Around this core manganese and iron oxides and hydroxides form more or less alternating concentric layers. However, these manganese nodules are product of immensely slow growth, the rate of which Pettersson tried to measure by radioactive methods.



54. Manganese nodules from the central Pacific. Photo of series of microscopic sections by the German Valdivia expedition. Note the radial growth around a nucleus.

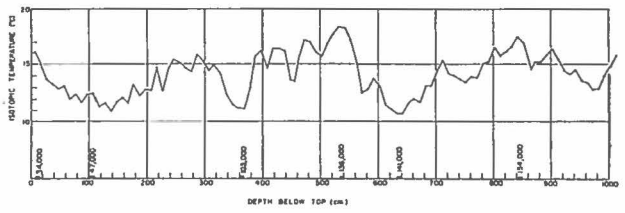
The Albatross brought also home a large harvest of manganese nodules (Fig. 55). Pettersson measured the rate of the radial growth of the manganese nodules by analyzing the radium present in thin concentric layers removed from manganese concretions. The result showed a growth rate of one millimeter per thousand years. This was the first known case where the disintegration of radium itself had been used for chronological purposes. As stated above, the migration of radium within the nodules due to recristallisations, made such measurements very uncertain. The rate of growth is now considered much less, i.e. 1–6, occasionally up to 300 millimeters per million years.

Usually the nodules lie partly buried in the sediment and growth occurs both below and above the sediment – seawater interface. If the rate of growth of the nodules is lower than the accumulation rate of the sediments, why do nodules not become buried? One explanation is that seismic vibration moves the con-

55. Professor Hans Pettersson with a harvest of manganese concretion from the North Atlantic floor.

cretions to and fro by the friction against the seawater. Experiments and theoretical calculations have shown that larger particles in that way move upwards causing an inverse graded bedding (Olausson and Uusitalo 1963). An earthquake larger than 4 on the Richter scale lasting for a few seconds has already this effect and the stronger the earthquake, the faster is the upward movement.

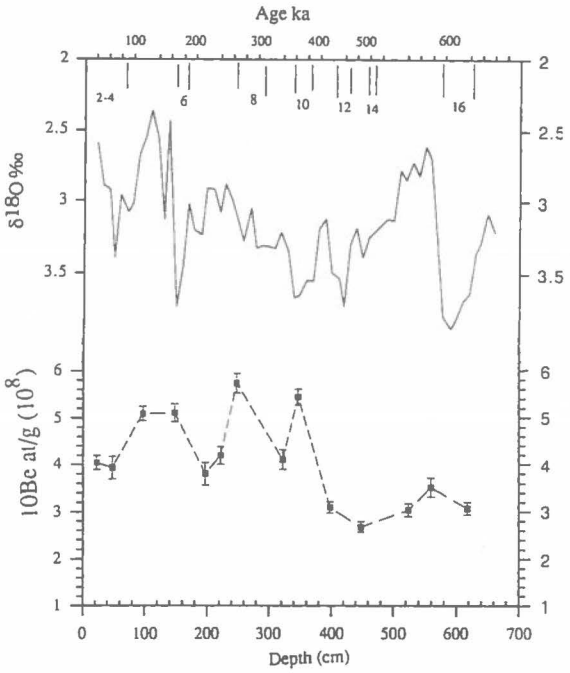
When Pettersson retired in June 1956 he left the Institute of Oceanography and later spent a year in Hawaii. Also Koczy left Gothenburg and went to the Marine Laboratory in Miami. That meant the end of the dating attempts in Gothenburg. Koczy and his Miami group took over these investigations and five years later an important paper appeared. Rosholt et al. (1961) published datings by Pa^{231}/Th^{230} of samples from Core 280 (Fig. 56) and two other cores. These datings are still considered as acceptable. G. Arrhenius, who had been the pioneer in introducing the C-14 datings to deep-sea carbonates (Arrhenius et al. 1951), also took up the development of the uranium-thorium dating methods of deep-sea sediments. Together with Goldberg, Koide, and Sackett (see e.g. Arrhenius and Goldberg 1955; Goldberg and Koide 1961) he obtained good dating results in La Jolla with this method.



56. Core 280 with Pa^{231}/Th^{230} dates and paleotemperatures. Redrawn from Emiliani, (1958) and Rosholt et al., (1961). The dates are: 8–16 cm = 34,000 B.P., 101–110 cm = 47,000, 361–367 cm = 103,000, 534–539 cm = 136,000, 633–639 cm = 141,000, and 843–849 cm = 154,000 B.P. The three upppermost dates could be too old.

Professor Peters of the Tata Institute for Fundamental Research in Bombay got samples from Albatross cores for dating with the radioactive element beryllium, ^{10}Be. This isotope is produced through atomic disintegration by cosmic rays in the upper atmosphere from where it ultimately settles on the ocean floor. His group studied the radiolarian ooze of Core 73 and found that the ^{10}Be concentration "indicated a sedimentation rate of about 5 mm/1,000 years" (Goel et al., 1957). Recent attempts to use this isotope for dating has also been made by Aldahan, Possnert & Olausson (1992), using samples from two Atlantic cores, 290 and 299 (Fig. 57). The distribution of ^{10}Be is not yet fully understood.

The C-14 dating method is reliable for the interval back to ca 40,000 B.P. Uranium – thorium datings by mass spectrometry are used for coral samples from the last ice age. K/Ar – datings of volcanic ash is used to date magnetic polarity change further back in time with results that may be applied to date deep-sea sediments. The cyclicity of the carbonate or ∂18-curves is assumed to be due to the so-called Milankovitch effect and thus tied to the chronology of the insolation curve. However, the validity of the Milankovich "dating", i.e. the periodicity of about 40,000 years remains to be proved. The deep-sea chronology beyond the ^{14}C-time scale has unfortunately a character of being a rubber-band chronology where the isotopic curve can be stressed or compressed at will.



57. Matching between ^{10}Be content of bulk samples and $\partial^{18}0$ of the *Cibicides wuellerstorfi* from Core 290. Preliminary isotopic stages are given. From Aldahan, Possnert and Olausson, (1992).

58. In August 1965 dr Arthur Swansson and I had the pleasure to visit Professor Hans Pettersson at his summer home, Bockholmen, close to the Bornö Hydrographical Station in the Gullmar Fjord, Sweden. While I presented him a festschrift, dedicated to him by his former students and associates on the occasion of his seventy-fifth birthday, Arthur used the camera. – Hans Pettersson, was born 26 August 1888 and he died on 25 January, 1966.

IX. NOTE ON THE INORGANIC GEOCHEMICAL STUDIES

About 16,000 samples were analyzed for the content of $CaCO_3$, C, N, P_2O_5 and those have been published by Arrhenius (1952) and Olausson (1960 a-d). These data have been discussed in Chapter 3. Berrit and Rotschi (1956) carried out analyzes such as the content of Fe_2O_3 and MnO_2; in many cores the Ni0 content was determined also. Landergren (1954, 1964) carried out a large number of spectrochemical analyses of bulk sediment samples. Since the results are difficult to summarize in short I can only state here that his data reflect the ages and the type of deposits analyzed.

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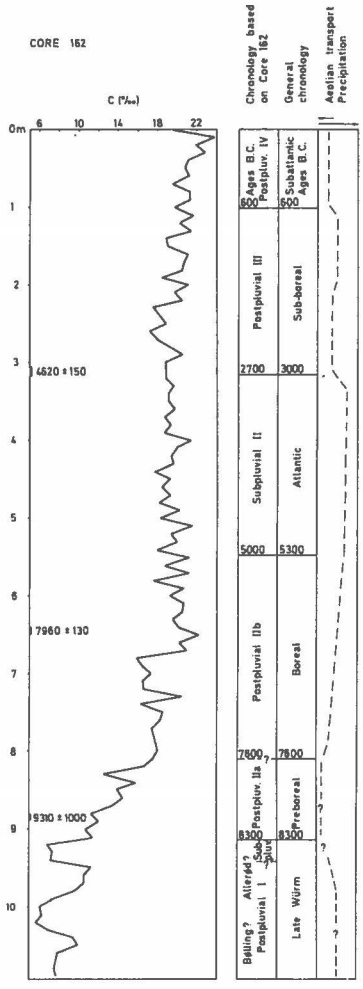
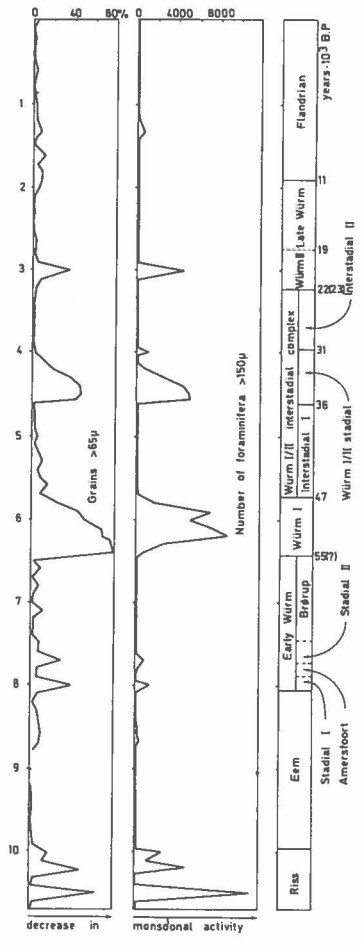
The extensive knowledge in the chemistry of sediments from different environments achieved during the studies of the Albatross cores was early used in tracing the man-made effect on sediments and water in our coastal areas. We had developed methods how changes in productivity, Eh and pH could be traced and interpreted. On this base we started monitoring research in Skagerrak-Kattegat during the mid-60's (summarized in Olausson, 1975; cfr. also Olausson and Cato, 1980) and training courses (see e.g. Olausson 1970 b). Our experiences then activated ICES to establish a Working Group on Marine Sediments in Relation to Pollution which still is active.

X. Late Quaternary changes of the Monsoons in the western Indian Ocean

The yearly fluctuation in the supply of eolian matter to the **Gulf of Aden** causes microstratification with an average thickness of 1 mm per microvarve. This varve stratigraphy has been correlated by C-14 time scale (Fig. 59). It was shown (Olausson and Olsson, 1969) that there is a reverse correlation between the content of eolian matter and the phosphate content in Core 162 from the Gulf of Aden. The phosphate curve could thus reflect the amount of precipitation in adjacent arid lands during the late Pleistocene. There is a marked drop in the precipitation during or after the Allerød subage. The Preboreal was the driest subage during the Holocene and a time of great wind erosion. A marked increase of the precipitation occurred during the Atlantic subage. The first millenium of the Subboreal subage was comparatively dry, whereafter (≈ 1800 B.C.) a slight increase of precipitation occurred. A new de-



59. The chemistry and stratigraphy of core 162 from the Gulf of Aden. ^{14}C ages are given in B.P. The curve of precipitation in adjacent arid zone (or the amount of eolian matter, transported to the Gulf of Aden) is a generalized phosphate curve.



60. Changes in the monsoonal activity during the Upper Pleistocene. Data from Core 156 (Olausson et. al., 1971).

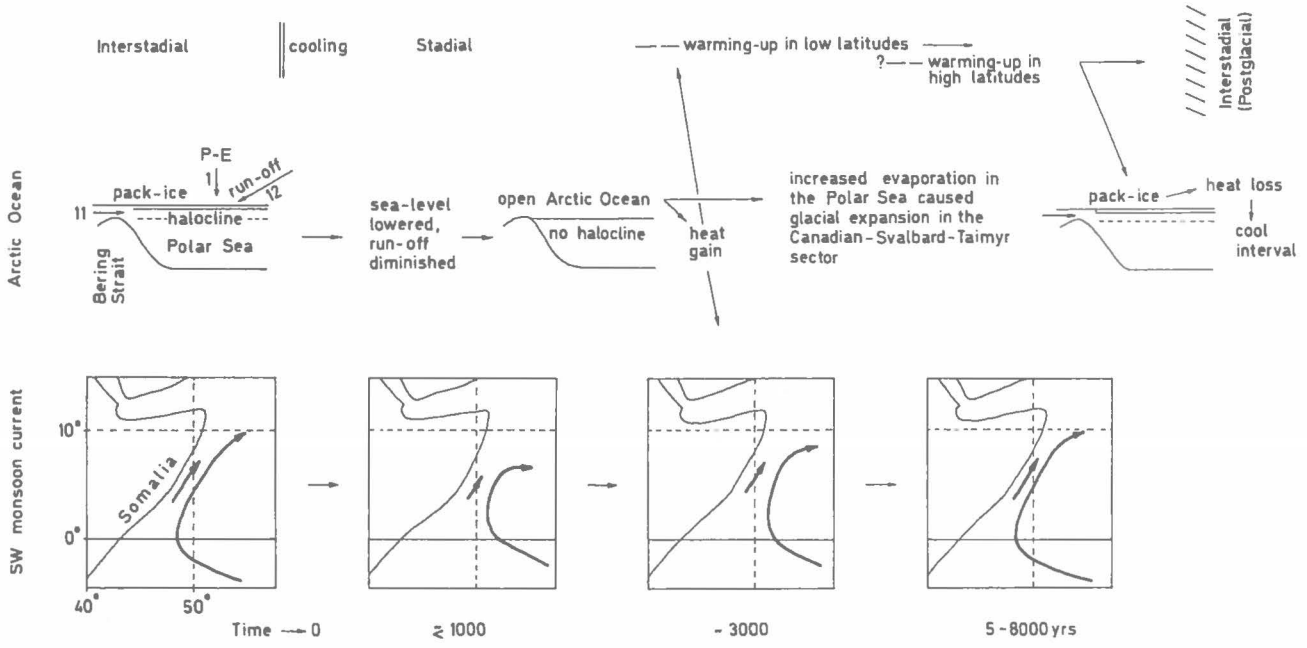
crease in rainfall gradually took place after 1200 and 700 B.C. Further details can be found in Fig. 59.

The sediments which cover the **Somali Basin** (the cores 155–158) consist mainly of coccoliths (Olausson 1960 b). Near the coast, below the Monsoon currents, and in the upwelling zone foraminiferal tests are frequent. Sediment cores from the basin consist of two types of carbonate oozes. Coccolith chalk ooze constitutes the main portion of the sediments. At certain horizons foraminiferal beds are found. The alteration between coccoliths and foraminiferal chalk oozes is explained in terms of paleoceanographic and paleclimatologic changes (Olausson et. al, 1971). It is thus found that changes in the paths of the Monsoon Currents caused the alteration in the lithology. The foraminiferal beds represent abnormal atmospheric and oceanic circulation regimes which prevailed during cool phases. Five such beds are recorded (Fig. 60 and 61).

I observed that in the Equatorial Indian Ocean during the isotopic stage 4 (= Weichselian I; ≈ 55,000–47,000 B.P.) the strongest deviation from the present circulation pattern was reached within the first 1,000 years of cooling; thereafter a gradual return to previous conditions occurred (Fig. 61). Furthermore, I observed that during a Weichselian I/II Stadial (≈ 36,000–33,000 (–31,000) B.P.) a rapid southward shift of the trade wind belt occurred within about the first 500 years of cooling, while the northward shift back to its present position happened gradually within some 3,000 years. Finally I noted that Weichselian II (≈ 22,000–18,000 B.P.) was a time of gradual cooling followed by a gradual warming of the Equatorial Indian Ocean.

The onset of a stadial, e.g. Weichselian I, apparently is not a question of either a simple, slow temperature decrease at higher latitutes or a global one. Instead we must think of a factor which is able to quickly change the atmospheric and oceanic circulation pattern. In addition, there may exist a factor in the Earth system which after some 1,000 years is able to increase the monsoonal activity. I believe that this involves a change of the Arctic Ocean from pack-ice covered into ice-free conditions which happens within the first 1,000 years of a stadial. This causes an appreciable reduction of the snow- and ice-covered areas in the northern hemisphere, i.e. the source regions of cool air masses (Fig. 61). This will reduce the meridional component of atmospheric transport and the monsoonal activity during northern summer will increase.

Reports of similar rapid climatic changes are given by e.g. Dansgaard et al. (1971, 1993) from Greenland ice core studies and by Van der Hammen et al. (1971) from pollen data from Macedonia.



61. Below: The location of the Southwest Monsoon Current during a stadial (Würm I = Early Weichselian). Time 0 represents present conditions (equalt to the Odderade interstadial). The strongest deviation from present circulatory pattern was reached within the first 1,000 years of cooling; thereafter a gradual change back occurred.
Above: Attempt to explain the cause of the rapid change in climate, traced in the Weichselian location of the South West Monsoon Current.

XI. Results from the Red Sea cores

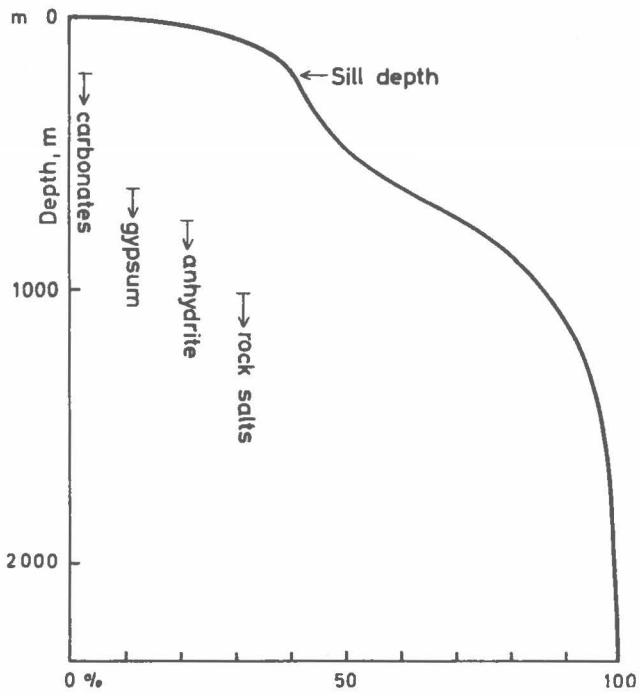
General comments

The two semi-enclosed seas, the Red Sea and the Mediterranean, are treated separately since their late Cenozoic development is partly different from that of the oceans.

The Red Sea is a subtropical semi-enclosed sea. The salinity is high owing to its small sill depth of 145 m, (Hanish Sill, according to the Royal Navy´s nautical charts no. 6 and no. 453) and associated limited water exchange, high evaporation (2 m per year), and lack of fresh water supply. The salinity is 37 o/oo at the entrance in the Strait of Bab el Mandeb, and it is then increasing northwards, reaching 41–42.5 per mil in the Gulfs of Aqaba and Suez. The surface inflow appears to be about 3 x 10^3 km^3 per year, nearly 1 km^3 evaporates, and thus approx. 2 x 10^3 km is returned back to the Indian Ocean. The surface waters of the upper 0– 60 m and the deep and bottom waters below about 200 m are homogenous, with an intermediate layer of complicated and variable structure in between. Inside the sill this intermediate water is about 24°C and 39.8 o/oo, but during its flow over the sill into the Indan Ocean, mixing reduces these values to 15°C and 36.5 o/oo. This outflowing water is readily identified in the Arabian Sea and reaches along the western side of the Indian 0cean of about 600 m as far as 25°S in the Mozambique Channel (Pickard and Emery, 1982, p. 212; Dietrich et al. 1988, plate 5.).

During glacial maximum the sea-level may have dropped to or below the sill depth. No surface inflow could then have compensated for the evaporation loss within the Red Sea.

In my contribution to the meeting of the SCOR Working Group no. 19 in Cambridge 1967, I made a theoretical account of the course of events during a complete evaporation of the Red Sea (Olausson 1971 a, p. 388–389; see also Fig. 62 redrawn from that paper). This account is here repeated shortly. Assuming an evaporation rate during glacials of 2 m per year and a cross-section area at the sill depth of 175 x 10^9 m^2, the water level of the Red Sea will fall below the sill depth when the inflow becomes less than some 350 km^3 per year or about 10 per cent of the present inflow (Olausson 1971 a). Then, most of the Red Sea area (90 % of the present area) would be dry after roughly 500 years, and after some 1100 years of isolation only some deep holes would contain brine as long as connate water was added from below. During the evaporation aragonite and dolomite were precipitated during the first phase. When salinity had reached to 3.35 times the normal value (about 117 per mil) gypsum would have started to precipitate. Theoretically this happened when the water level stood at about –600 meters. When the water volume was reduced to about one fifth compared to normal sea water (i.e. the water level stood at about –800 meters) anhydrite would have been the stable phase. When the salinity of the residual water was about ten times the salinity of normal sea water (i.e. at a water level of about –1000 meters in the median valley rock salts would also have been precipitated.



62. Hypsographic curve of the Red Sea. Levels are marked below which indicated salts could have been precipitated during a complete evaporation.

When the sea level would be the same as the present sill depth, an inflow of ca. 500 km^3 sea water per year (≈1.7 x10^5 m^3 per sec.) could have kept both the water level and the salinity more or less constant. A further decrease in the inflow would cause a rapid drop in the water level. An inflow of about 10^{11} m^3 per year (≈ ca

400 m^3 per sec.) would have compensated for the evaporation loss in the median valley.

If the sea-level had temporarily dropped below the sill depth of the Hanish Sill (13°40'N) then a subsequent (postglacial) transgression could have filled the basin within a few decades or so depending on the intensity of the melting of the northern hemisphere continental ice sheets. During this transgression rock salts, precipitated below –1000 meters, and anhydrite and gypsum precipitated below 600–800 meters, would have been dissolved as soon as the salinity was less than the aformentioned concentrations.

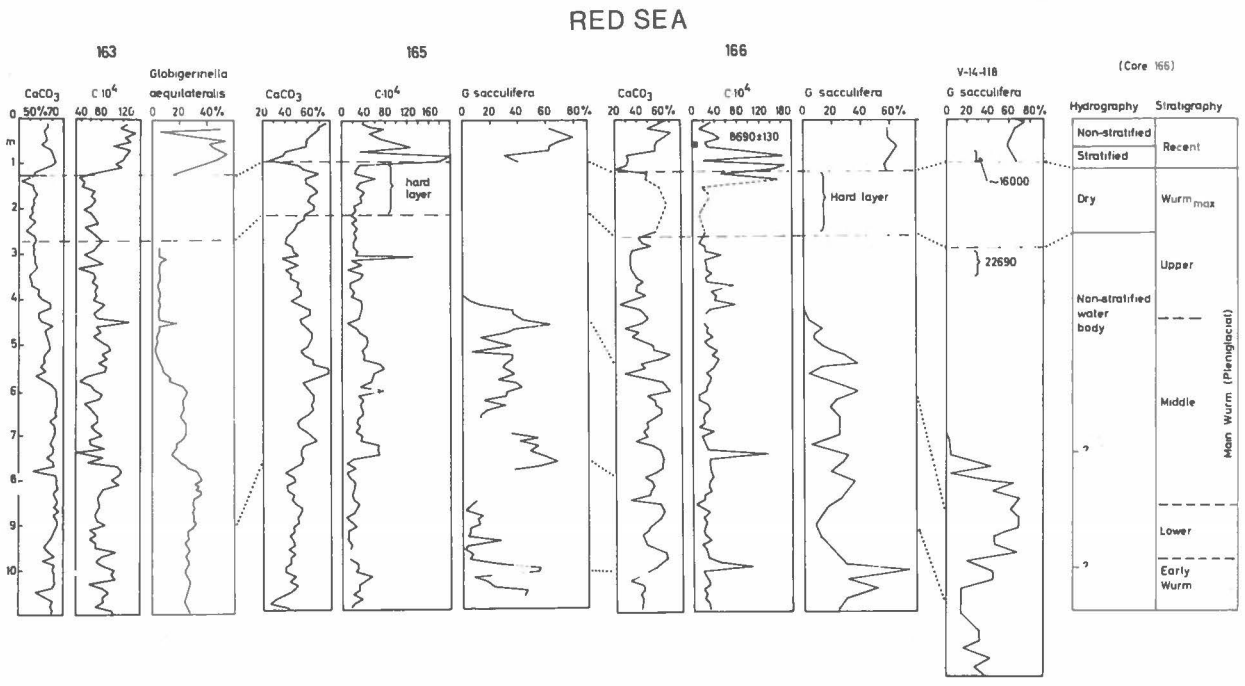
The Stratigraphy of Red Sea Cores

The Red Sea cores no. 163–169 collected by the Albatross Expedition (Fig. 14), were described by Olausson (1960 c) in which paper also the chemical data from the cores were given. I discussed the cores in more details later on (Olausson 1969 and particularly 1971a). Here I will summarize the stratigraphy from the papers.

The cores can be subdivided into three sections indicative of three different environmental stages.

Section 1. The basal part of Cores 165 and 166, up to about 250 cm, consists of foraminiferal chalk and marl muds. This section is formed during the early – middle Weichselian (Fig. 63) in a comparatively cool, non-restricted Red Sea.

Section 2. Between 97 and 204 cm in Core 165 and at 113–260 cm in Core 166 (below 65 cm in Core 169 and at least 199–221 cm in Core 168) there are beds with large irregular lumps of calcareous layers interbedded with marl mud. The hard lumps sometimes have a smooth surface filled with pteropods, the other side is pitted. The matrix is a felt-like intergrowth of aragonite aggregates containing particularly pteropods, solitary foraminifers, a few fragments of echinoids , etc. The orientation of the fragments of calcareous layers is more or less random, suggesting that it was formed by a crushed and reworked layers into which marl mud has been mixed. The bulk mineral is aragonite, with magnesium-calcite (ca 15 mol-percent Mg) being the second important mineral. The hard bed sequence (*sabhka*) is more frequent in our cores from below coring depths of 800 m than from shallower water. In the cores from the valley the section 2 is rich in carbonate, but poor in phosphate and planktic foraminifers. Pteropods are present, however. Judging by the C-14 datings (Fig.



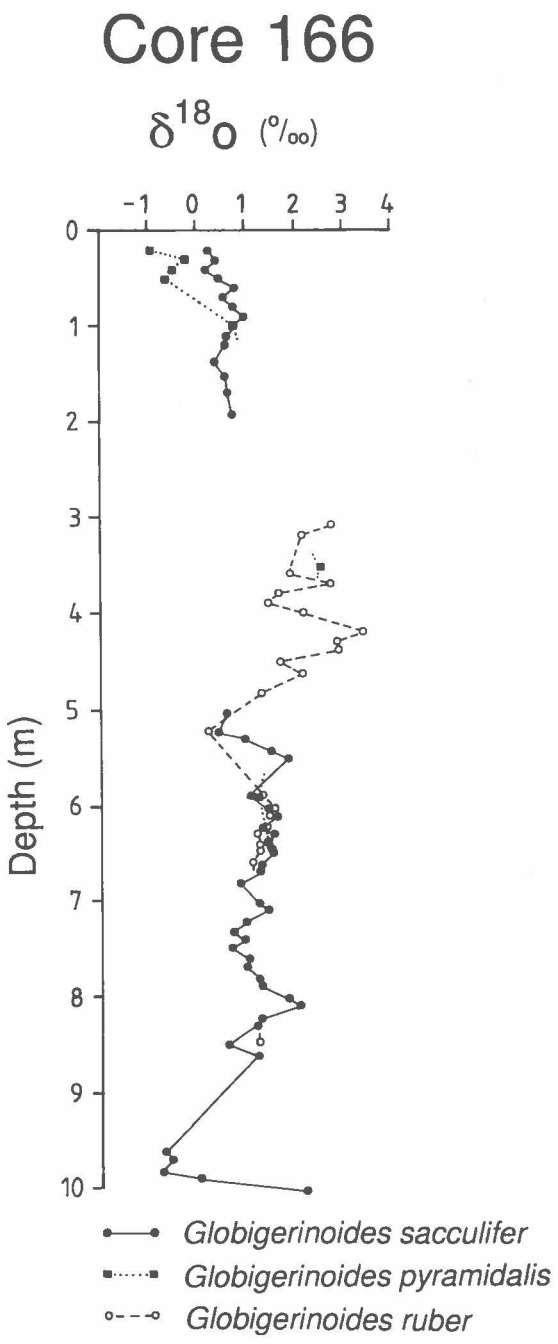
63. Stratigraphy of four Red Sea cores. The C-14 ages are given in years B.P.
The hard layer sequence (sabkha or dissolution breccia) is found over large areas of the Red Sea floor. Judging by cores 165 and 166, collected from a depth larger than 1,000 m, the axial valley was stagnant (high carbon content) and the water may have been slighly acid (low $CaCO_3$). Above 1000 m depth (Core 163) the water has been normal during the Late Weichselian (= Late Würm) and Holocene (= Recent). The stagnation of the water in the axial valley lasted to approx. 8,000 B.P. (carbon decrease).
As appears from Fig. 62 rock salt may have been precipitated below 1,000 m.

63) the hard bed sequence was formed somewhere between 16,000 and 22,000 B.P., i.e. during the last maximum of the Weichselian ice age. The foraminifer-barren hard bed can be understood if we envisage that the sediment surface layer was cemented together by aragonite during an evaporative phase. Taking into account the porous nature and angular shapes of the lumps, and considering the wide-spread occurrence of the hard bed, it may constitute a sediment bed reworked by wave action and mixed with some other deposits.

The hard layer (section 2), according to my interpretation (Olausson 1971 a, 1969, and 1970), is a result of high-salinity phase of the Red Sea during the late Weichselian ice age. To what extent the basin was dry 18,000 years ago is hard to determine. Even if we know the present sill depth we do not know the sill depth at that time since erosion and isostatic adjustment may have altered its present height. We do not know the sea-level stand 18,000 B.P. ; I often use the conservative figure of 100±20 m. My isotopic data (Fig. 64) show that the $\partial^{18}0_{calcite}$ increases by about 2 per mil in the uppermost part of section 1. From the foraminifer-barren section 2 I have no isotopic data except for precipitated aragonite (+ 7 o/oo). This figure has also been found by Deuser and Degens (1969). The scarcity of foraminifers suggests that salinity had become too high for these planktic organisms. According to Hemleben´s laboratory studies (Hemleben 1986) the upper salinity limit for *G. sacculifer* is approx. 45 o/oo and for *G. ruber* approx. 47 o/oo. Aforementioned ∂-figures from Core 166 (Fig. 64) point to a salinity of about 50 o/oo during the earliest phase of stage 2, which explains the absence of foraminifers. The aragonite samples indicate salinities of > 60 o/oo. That is all that we know today. Later studies of other Red Sea cores have only verified the data from the Albatross but usually did not refer to these earlier results (see e.g. Locke and Thunell 1988; Brachert and Dullo 1990).

Section 3, younger than approx. 16,000 B.P. Above the hard bed sequence there is a marked increase in foraminifers (particularly *Globigerinoides sacculifer*) and pteropods (e.g. *Limacina inflata*).

The foramineral and chemical data and stratigraphy for the Red Sea cores 163 (coring depth: 300 m), 165 (1158 m) and 166 (1283 m) are summarized in Fig. 63 (redrawn from Olausson 1971 a, Fig. 29.16). Core 163 was raised from the area where carbonates could be assumed to have been precipitated



64. $\partial^{18}0$ of Core 166 from the Red Sea (17° 56´N, 39°57´E; depth 1283 m). All values are relative to PDB. Three species are analyzed: *Globigerinoides sacculifer, G, ruber, and G. pyramidalis* (van den Broeck). The last is a subspecies of *G. ruber* (cfr. Parker, 1957, p. 279)

during an evaporation phase. The other two core from sites where rock salts could have been precipitated and where high salinity conditions could have prevailed during a period when the inflow of Indian Ocean water was considerably reduced or totally ceased (see calculations above). The boundary between the sections 2 and 3 is recognized by an increase in organic matter, phosphate, and calcium carbonate (foraminifers). In Cores 165 and 166 there are

increases in carbon and nitrogen but a decrease in calcium carbonate.

The late Weichselian – early Holocene period, approx. 16,000–8,000 B.P. was characterized by reduced ventilation of the water in the axial valley. In order to date this event a C-14 dating was carried out on a sample from Core 166, 52.5–58.5 cm (Olausson and Olsson, 1969) which gave the age of 8,690 ± 130 B.P. This age represents the end of the stagnant phase of the axial basin coinciding with a drop of the carbon content from about 2 % to 0.03 %. The last figure is a normal percentage for a non-restricted Red Sea (see Fig. 63). The high carbon content exclusively found below a depth of 1, 000 m suggests that an oxygen deficiency prevailed in the axial valley during these approx. 8,000 years. This implies that the vertical circulation in the Red Sea below approx. –1,000 m was reduced or ceased due to a salinity stratification with a strong halocline at aforementioned depth.

A calculation about the time required to smooth out such large differences in salinity only by diffusion points to approx. 8,000 years as well (I. Rosenquist, pers. com.). Whether salts have been added also from the Miocene salt deposits further down in the rift or trapped high-salinity water originated from the high-salinity phase only is not known. – Hydrographic change at the Pleistocene/Holocene boundary is not clearly distinguished in our Red Sea cores; however, the carbonate increase in Core 163 could indicate this boundary.

The brines found in the Red Sea during 60´s are mentioned in Chapter IV:4.

The carbonate minimum in the cores from the valley in connection with the stagnations (not found above the –1,000 m!) was interpreted as due to a decreased pH in the supersaline water of the valley axis (Olausson, 1971 a, p. 392).

Foraminifers such as *G. sacculifer* have a tendency to increase in frequency from south to north, while other species such as *G. glutinata* and *G. ruber*, decrease in relative frequency. This may be be due to the decrease in temperature towards the north or/and the fact hat *G. sacculifer* has higher tolerance for high salinity than the other species (l. op. cit.).

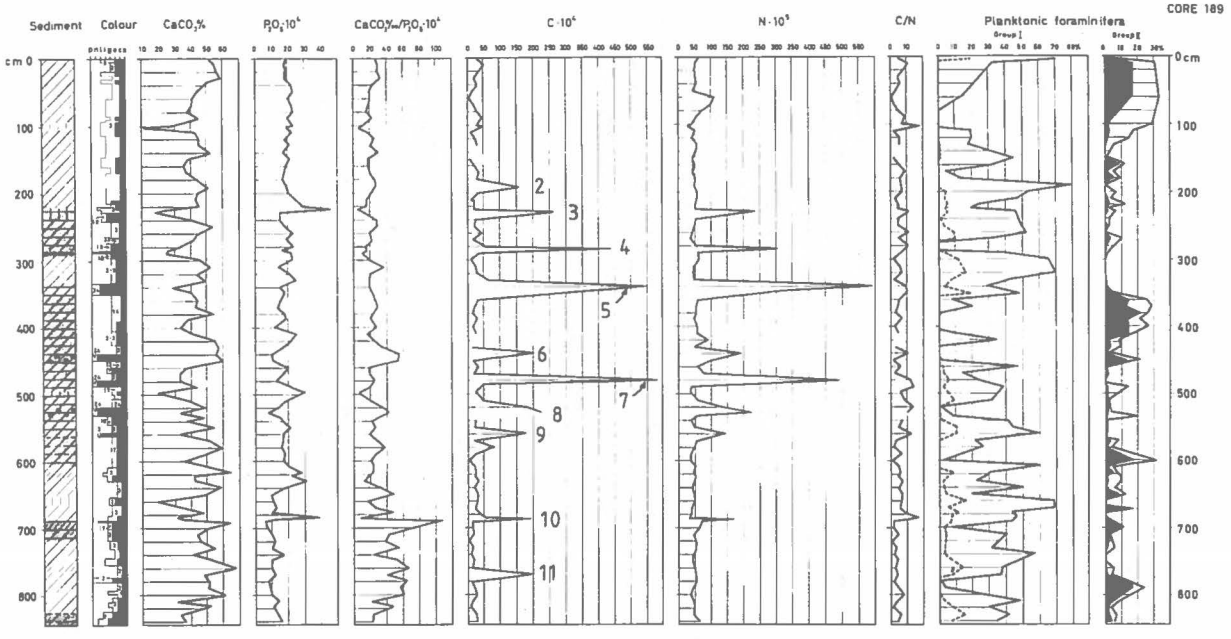
XII. The Mediterranean and its last twelve stagnations

The most important studies of the Mediterranean cores 176–211 (see map Fig. 12) are those published in vol. VIII of the Reports from the Albatross Expedition. These are sediment-petrographic studies of the western Mediterranean cores by Erik Norin (1958), S. Duplaix (1958), and K. Gösta Eriksson (1965), studies of the foraminifers in the western basin by Ruth Todd (1958) the eastern basin by Frances L. Parker (1958), and the paleoceanographic studies of the eastern Mediterranean by myself (Olausson 1960, 1961). The foraminifers have already been dealt with in VI:4 and the abnormal fauna will be treated in connection with the sapropelitic muds.

The cause of the late Quaternary stagnations of the eastern Mediterranean

The sediment cores collected from the eastern Mediterranean showed alternating layers of white calcareous oozes and dark (black) mud with a higher amount of organic matter (Fig. 65). The dark bands suggested that different (anoxic?) bottom conditions had prevailed during certain intervals back in time.

Bradley (1938) had theorized about the glacial Mediterranean. He assumed that there was a blanket of a low-salinity surface water during pluvials giving



65. The chemistry and lithology of Core 189 from the Levant (33° 54´N., 28° 29´E ; 2664 m depth). Note the sapropelitic layers no. 2–11. The latest sapropelitic bed, no. 1, is lacking in this core. The carbon content are highest in the beds 5 and 7, which stagnations may have been the strongest among the post-Elsterian stagnations.
The section down to 3 m is Weichselian in age. The sapropel bed no. 5 is formed during early Eemian. From 3.6 to 6.6 m the section is formed during the Saalian (isotopic stages 10-6), and below 6.6 m during the Holsteinian interval (stages >10).
The foraminiferal group I was considered as "warm" and the group II as "cold" according to Parker (1959, Fig. 6). The frequency of *Neogloboquadrina (Globigerina) pachyderma* is given as the black part at the base of curve II. – From Olausson, (1960 c).

rise to a change of the current directions in the Strait of Gibraltar. The oxygenated Atlantic water would have had free access to the western basin of the Mediterranean and the deep water there would have continued to be at least partially supplied with oxygen (l. op. cit. 377). Young sapropelitic muds are indeed lacking there but occur in older sequences (core 201, 1015–1018 cm with a tentative age at the isotopic stage 6/5 (= Saalian/Eemian transition). Under such conditions "there appears to be no way in which significantly large volumes of the oxygen-bearing saline water in the western basin could move across the sill at the Sicily Straits so as to cause circulation and afford an oxygen supply in the deep saline water of the eastern basin." – Under these conditions "the stagnant water becomes poisoned with hydrogen sulphide which kills off all the bottom fauna. With no bottom dwelling organisms to disturb the sediment and aid in the destruction of the sediment accumulates as successive pairs of thin laminae, one of which is rich in organic matter" (Bradley op. cit., p. 377) "In short, varved sediments would have accumulated as they are accumulating to-day in the deep water of the Black Sea" (l. op. cit. p. 378). Such sedimentary conditions as predicted have been recorded in the cores as will be seen further below – but they occur during other time intervals.

Kullenberg (1952) studied the cores 187, 189, 190, and 198 from the eastern basin (Stations 274, 276, 278, and 287). and published their carbon and water content, the salinity of the interstitial waters, and the distribution of a few warm and cool water indicating foraminifers. He explained the seven black layers he found as due to reducing conditions. "Stagnation of the Mediterranean can be brought about by increased precipitation and run-off, and decreased evaporation, which should lower the salinity of the surface water" (op. cit. p. 32).

When I studied elementary geology in 1948–1950 I read in e.g. Flint (1949) that pluvial ages were contemporaneous with glacial ages – "that the expanded lakes and the ice sheets reached a maximum at the same time. – The growth of the former ice sheets should have provided the atmospheric mechanism required to produce the pluvial phenomena. The case is not entirely closed, but the probability is of a high order" (op. cit. p. 469). A similar conclusion was also reported by e.g. E. Nilsson (1938). when dealing with the pluvial lakes in East Africa.

Having read Kullenberg´s and Parker´s results, I became very interested in finding the origin of the black layers. I defined these as mud with a carbon content in excess of 2 %, and I described this sediment as **sapropel** (sapropelitic mud), a name later accepted by other scientists. Judging by the foraminiferal analysis I found that the uppermost sapropel occurred **after** the Weichselian maximum, at about the Pleistocene/Holocene transition. It was then necessary to get C-14 dates of the uppermost sapropelitic layers. My colleague Ingrid Olsson carried out four datings (Olsson 1959) which indicated that the time of sapropel formation was not as Bradley or Kullenberg had thought: the glacial (pluvial) times. Sapropel formation occurred instead during the warming phases, i.e. at the time of the transition from the late glacial (sub-)ages to the following early interglacials (interstadials). The last stagnant phase started approx. at the Allerød/ Younger Dryas transition (≈11,000 B.P.), when the sea-level had raised to about – 40 m (=the present sill depth of the Bosporus) and I found that the euxinic phase ended approx. during the 6th millenium B.C. (Olausson 1961 a, p.354). Later studies date also the last euxinic phase at 11,000–7,000 C-14 yrs B.P. – The idea of a pluvial during ice ages was apparently wrong. I later learned that the pluvial periods belonged to the climatic optimum during interglacials (Olausson and Olsson, 1968, Olausson 1971 b; see also Fig. 69).

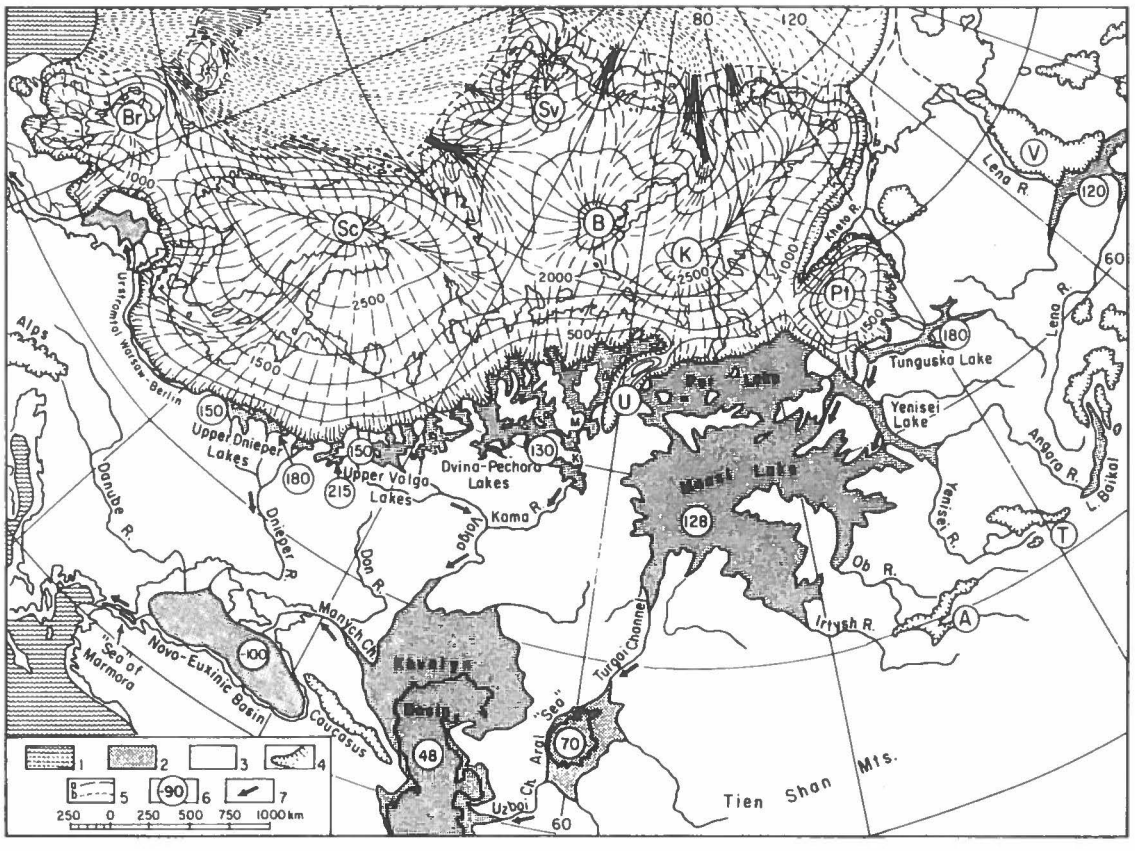
The actual negative water balance in the eastern Mediterranean (EM) is a result of excess evaporation (E) over precipitation (P), plus runoff. The present runoff to this basin from the Nile is ca. 0.001 Sv (1 Sv = Sverdrup = $10^6 m^3$ per sec.) and the Po and other rivers give 0.0015. Runoff from all rivers is thus ca. 0.0025 Sv. Evaporation from the EM is today 1.6 m per year, the present precipitation is 0.6 m/yr and E-P is thus ca. 1 m (= 0.06 Sv). Evaporation should have been lower during the anoxic phases while P would have been higher and E-P could have been around 0.6 m or ca. 0.03 Sv. In order to change the water balance from negative to positive we need at least one additional fresh water source of this magnitude. This quantity is ca. 30 times larger than the water discharge from the Nile. We can thus exclude the Nile for having caused the stagnation. In addition, the last pluvial maximum began when the stagnation ended. Besides, a sudden lowering of the $\partial^{18}0$ in foraminifers by 2–3 per mil would, if caused by the Nile (= strong monsoons), corresponds to a salinity drop of the surface water of the EM to at least 25 per mil. Such a low sur-

face salinity is unlikely because of several biological reasons. Therefore interpretations of the cause of the stagnations as given by Rossignol-Strick et al. (1982) appear to be unrealistic. However, an increase in the river water supply to the EM up to 0.005 Sv during the anoxic periods is probable. A temperature rise from the glacial maximum to the interglacial optimum by 4°C (from ca. 8 to ca. 12°C), which is a realistic order of magnitude, must be compensated by a salinity increase by 0.8 per mil, if the density of the water is to remain the same. However, the warming of the water was probably a slow process and only 1°C is taken as likely during the early euxinic periods. Therefore, the density change due to climatic amelioration was a minor factor, being equivalent to a fresh water supply of approx. 0.008 Sv only (see also Olausson 1971 a, Fig. 12).

The present inflow from the Black Sea (BS) is ca. 0.011 Sv with a salinity of ca. 16 per mil and the outflow from the EM to the BS is 0.005 Sv. The fresh water discharge from the BS is thus 0.006 Sv. i.e. twice the supply by all rivers to the EM.

During intense deglacial times the BS must have received much more water than it does today, not only from the Danube (presently 0.006 Sv) and other rivers, but probably also from the Caspian/Aral Sea (Fig. 66). The section of the Russia lying west of ca. 110°E was drained to the south during glacial times, which means that the Black Sea/Caspian Sea would have received the water now discharged to the north by the rivers from Dvina to Yenissey, today ca. 0.03 Sv. As long as there was a remaining ice sheet in northern Russia (part of it) this water was discharged to the oceans via the Black Sea. Furthermore, meltwater from the disintegrating ice sheet in the north as well as from the ice caps in adjacent mountain ranges was also discharged to the Mediterranean via the Black Sea.

Let us make an attempt to calculate the fresh water flux from the BS to the EM during the last anoxic period of the latter basin (11,000–7,000 B.P.). The present fresh water discharge of 0.006 Sv is taken as the minimum value. A maximum value could be this



66. Last Eurasian Ice sheet and related proglacial drainage system. (1) Glacier-free deep-sea; (2) ice-dammed and other inland freshwater basins; (3) ice-free land; (4) boundaries of ice sheets and mountain glacier complex; (5) lines of ice flow on (a) terrestrial and grounded marine parts, and (b) floating parts of the ice sheet; (6) surface of inland basins, m; (7) directions of stream flow in channels of the drainage system. (M) Mylva Channel, (K) Keltma Channel. Ice domes and mountain glacier complexes: (Br) British, (Sc) Scandinavian, (B) Barents, (Sv) Svalbard, (K) Kara, (U) Uralian, (Pt) of Putorana Plateau, (V) of Verkhoyansk Mountains, (T) of Tuva-Sayan Mountains, (A) of Altai Mountains. From Grosswald, (1980), with permission from the Copyright Clearance Center, Inc.

volume plus all waters presently drained to the north by aforementioned Russian rivers, i.e. 0.03 Sv. Thus, the fresh water discharge from the BS to the EM would have been somewhere between 0.006 and 0.036 Sv during the (last) anoxic phase.

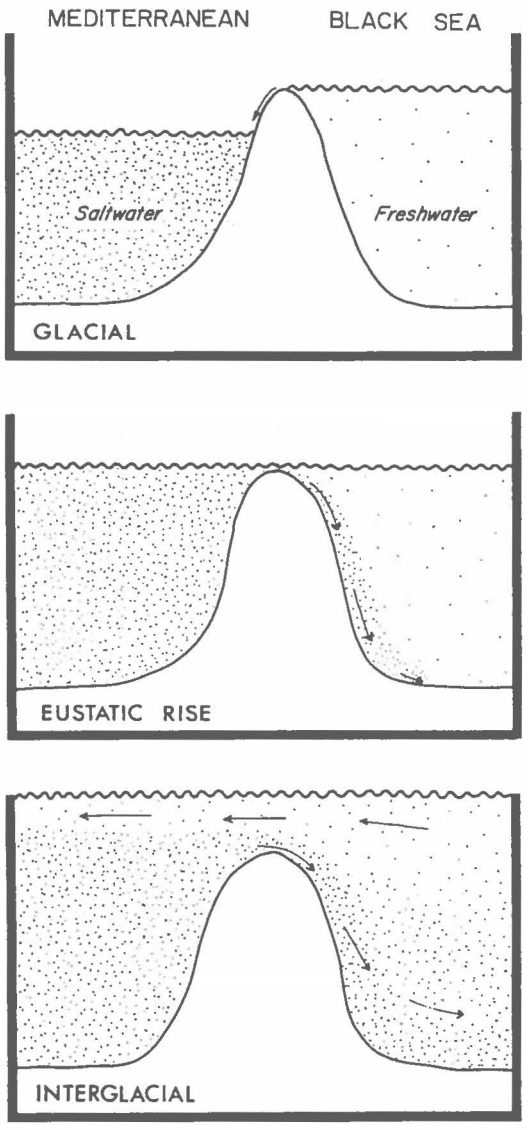
Judging by Degens et al. (1986, Fig. 3) the sediment accumulation rate of in the BS was ca. 100 cm per ka at the time when the anoxic phase began in the EM and it decreased to ca. 10 cm per ka during the climatic optimum when the EM became an oxidized basin again. There is generally a linear relation between runoff and transport of suspended inorganic material in natural rivers. Therefore, we can correlate the high accumulation rate with the 0.036 Sv water flux. Then 0.018 Sv could be an approx. figure of the fresh water flux from the BS to the EM for the time in question.

During the last ice age the connection between BS and EM was cut off and the BS basin was filled with fresh (glacial) water. During postglacial warming the EM water transgressed into the BS and about 0.5 x $10^6 km^3$ of the stored fresh water poured into the EM (Fig. 67). The present renewal time of the BS is ca. 2,500 years and over such a long period of time the average fresh water transport may have been up to 0.006 Sv (0.004 Sv if exported over a period of 4,000 years).

A fresh water budget of the eastern Mediterranean during the last anoxic phase may be summarized in the following tentative table [all units are in Sv (Sverdrup; 1 Sv = $10^6 m^3$/sec)]:

Glacial melting and river discharge via the Black Sea	0.018
River discharge direct to the eastern Mediterranean	0.005
Warming of the EM (equivalent)	0.008
Input from BS stored glacial water (flushing; Fig. 67)	0.006
Total input of fresh water to the EM	**0.037**
Calculated input required for causing anoxic conditions in the EM:	0.030

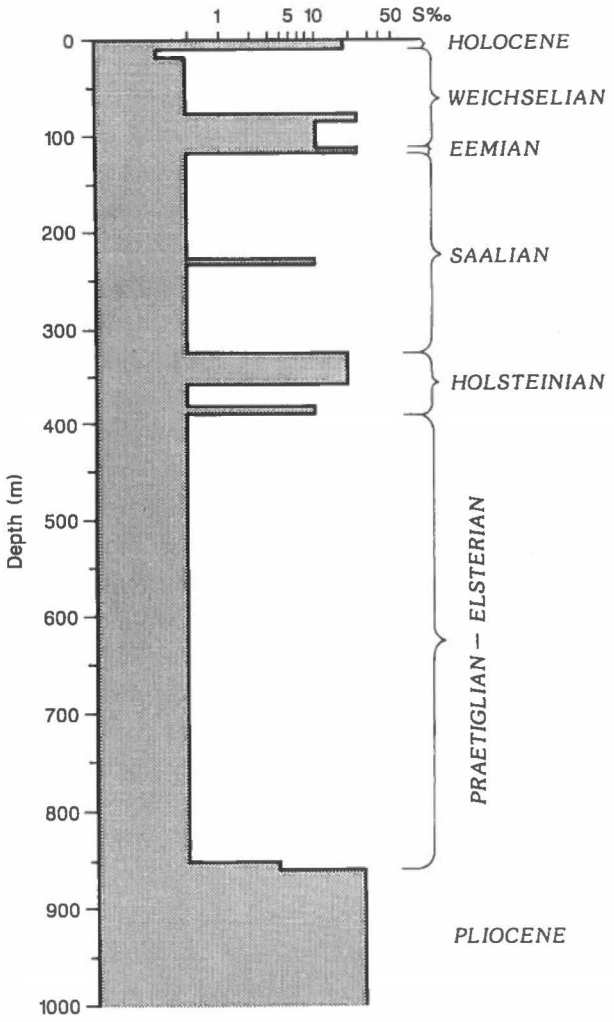
From this preliminary budget it is evident that fresh water input from the Black Sea is the main cause of the anoxic phases of the eastern Mediterraean. It constitutes ca. 85 % of the excess fresh water supplied to the EM. **Thus, fresh water outflow from the Black Sea triggered by postglacial sea water inflow into the Black Sea may be the key mechanism controlling stagnation in the eastern Mediterranean.**



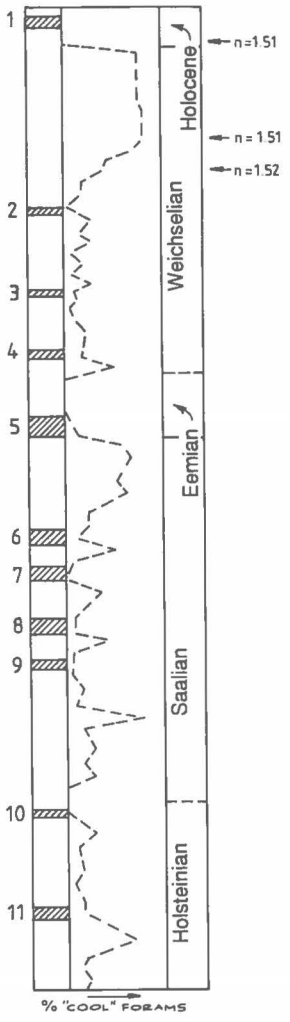
67. Stages of communication between the Mediterranean and the Black Sea. Low eustatic sea stands in the Mediterranean lead to isolation of the latter sea and its freshening. Subsequent rise of sea level allows denser marine waters to invade across the Bosphorus sill as an underflow, flushing out the former fresher water of the Black Sea, contributing to a brackish lens of the eastern Mediterranean (redrawn from Cita et al., 1973, Fig. 4).

On the 12 late Pleistocene stagnations

It appears from the above that fresh water outflow from the Black Sea (BS) is one of the most important causes of stagnation of the eastern Mediterranean (EM) (Olausson, 1961). Judging by Fig. 68 (redrawn from Schrader 1978), the BS was a marine-brackish sea during the last three interglacial ages (the Holsteinian, Eemian, and Holocene) while it formed a

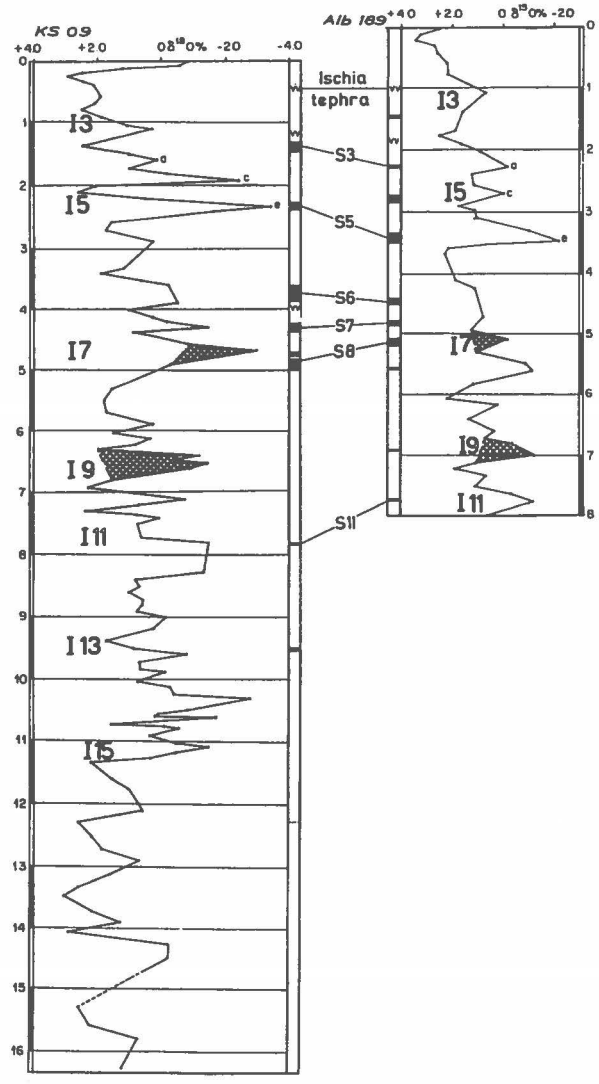


68. Generalized interpretation of environmental conditions in the Black Sea at diatomaceous interval and intermediate units in o/oo salinity (according to Schrader, 1978, Fig. 29) and the tentative Pleistocene stratigraphy by Olausson, (1965, 1982, 1988, and 1991 b). A similar result can also be reached from data by Mikhailescu et al., (1994). – Saalian = isotopic stages 6–10 and the Holsteinian marine phases may have occurred during the isotopic stages 11 and 13.



69. The figure show the distribution of sapropelitic beds no. 1–11 in the eastern Mediterranean core 189 and the Holocene section of core 194. The percentage of "cool" foraminifers according to Parker (1957) and the time-stratigraphy are given. The volcanic ash horizons are indicated. Note the existence of vitric ash with n = 1.51 at two different horizons (ages). Weichselian (isotopic stages 2–5 d) and Riss = Saalian (isotopic stages 6–10). The Holsteinian = isotopic stage 11.

fresh water lake during the early Pleistocene. A similar conclusion can also be drawn from the data published by Mikhailescu et. al. 1994. From this evidence it can be concluded that the sapropelitic formation in the EM should have been restricted to the post-Elsterian ages (from the isotopic stage13 or 15). This is also the case. The lowermost sapropelitic layer in the Albatross cores was S-11 (numbers according to McCoy 1974), which occurs during the isotopic stage 11(≈ Holsteinian s.s.; Fig. 69–70; cfr. Olausson, 1961 a). A still older sapropelitic layer, S-12, is recognized in Core KS09 (Fig. 70; Cita et al. 1977). This layer is somewhat older than the extinction of *Pseudoemiliania lacunosa*, but, thus, also within the Holsteinian complex, probably in the isotopic stage 13. Below this S-12 in Core KS09 there is ca. 7 m of sediment devoid of sapropelitic layers, accumulated during times of more stable hydrography (Fig. 70). Site 376 (Kidd et al., 1978,p. 439) contains a series of sapropelitic layers down to around ca. 13 m, corresponding to S-1 to S-12. Below 13 m and down to the Plio-Pleistocene boundary at ca. 39 m, there are only two sapropelitic layers recorded presumably formed during the early (-middle) Pleistocene. Therefore, the evidence from the deep-sea drillings supports the conclusion that the major cause of the late Quaternary



70. Correlation of Core KS09 and the Albatross core 189. Isotopic data obtained for *Globigerinoides ruber* are from Emiliani, (1955 b) and Vergnaud-Grazzini, (1977). Notice the sapropel bed no. 12 at 9.5 m and its absence further below in core KS09. Redrawn from Cita et al., (1977) with permission from Copyright Clearance Center, Inc.

stagnations of the EM was Black Sea water, as stated earlier (Olausson 1961).

The development of anoxic conditions and formation of sapropelitic muds in the EM did not only occur during rapid climatic amelioration associated with early postglacial warming as S-1 (early Holocene) and S-5 with early Eemian (isotopic stage 5 e), but also in connection with interstadials (S-2 to the Hengelo-Denekamp interval, S-4 to the Brörup interstadial (isotopic substage 5 c). S-6 to S-9 occurred during interstadials in the isotopic stage interval 6–10 (= Saalian ice age). S-10 to S-12 occurred within the Holsteinian interglacial complex i.e. in connection with early parts of the isotopic stages 11, 13 (and 15).

Thunell et al. (1983 a, 1983 b) stated that only S-6 is associated with a glacial stage. This is based on the correlation of S-6 to the isotopic paleotemperature curve from the Pacific, and on the assumption that the Saalian ice age is one single cold phase. The isotopic curves from the equatorial Pacific indicate an (over-) simplified climatic curve in which details in the complicated climatic development can not be resolved or may be overestimated. S-6 appears in a section where e.g. the Albatross core 194 indicates a change from cool to warmer conditions (see e.g. Olausson 1971 a, Fig. 3) and the $\partial^{18}0$ of *Globigerinoides ruber* from Core 15-PC (Parisi 1987) reaches the low isotopic value of – 3.57 per mil, which indiates a large supply of meltwater. Low $\partial^{18}0$ values are also found in samples of the same age from other areas of the EM [e.g. the Albatross Core 195 (unpublished data) and M76 (Lunkad 1981)].

ISOTOPIC ANALYSES

Emiliani (1955 b) studied the $^{18}0/^{16}0$- ratios of *Globigerinoides ruber*, *Globorotalia inflata* and *Globoquadrina dutertrei* of Core 189 (Fig. 70). I reinterpreted his isotopic curves ten years later (Olausson 1965) showing that the isotopic values were strongly influenced by the changes in the hydrography and by the meltwater supply from the Black Sea.

The amplitude of the ∂-values of Core 189 is 5.6 per mil which is about twice as much as in the open oceans. There is an abnormal peak at particularly S-5, where $\partial^{18}0$ dropped from + 0.45 (oxidized ooze from the latest Saalian) to –2.18 (in the sapropel) and then up to +1.05 per mil in the overlying (early Eemian) ooze. This period of stagnation is the most extreme during the upper Pleistocene. I calculated (l. op. cit. p. 230) that the surface salinity of the EM dropped down to ca. 33 o/oo, i.e. below that of the North Atlantic. Since S-5 is also found in core 201 from the western Mediterranean (see further below) it can be concluded that the currents in the Strait of Gibraltar were reversed (for the last time). A reversal of the currents across the Scotland – Greenland Ridge occurred at the same time due to the huge meltwater discharges (Olausson 1983, Fig. 1). S-8 represents another strong stagnations and can probably be correlated with the Saale 1/Treene- transition (isotopic stage 10/9). *Globigerinoides ruber* lives mainly in the uppermost ca. 50 m of the surface water layer while other species normally prefer a deeper habitat. In the layers from the euxinic phases, *G. ruber* often has 2–3 per mil

lower (more negative) ∂-values than *G. bulloides* (in S-5), in S-8 up to 4 per mil (Olausson 1991 a). This suggests that the low-saline water (with low $\partial^{18}0_{\text{sea water}}$) was mainly confined to the uppermost ca. 50 m, the larger the isotopic difference the stronger was the stagnation.

THE SALINITY OF INTERSTITIAL WATER

Kullenberg´s attempt (1952) to determine the past salinities through determination of the chlorine in core samples was not successful. He obtained large variations in the salinities, high values in the sapropel and lower in the carbonate oozes. The maxima coincide with the carbon maxima. He stated that "it is deemed unprobable that the occurrence of salinity maxima in layers rich in organic matter is due to adsorption of chlorine in these layers, though the possibility that such is the case can not be definitely ruled out" (op. cit. p. 36). – The answer to the problem is probably ion filtration during the early diagenetic compaction of the sediments. Water and non-charged matter pass through the overlying layers where particularly the organic matter may act as a ion trap. Thus, the salinity of the interstitial water does not mirror the salinity of the past bottom water. It fluctuates with the content of organic matter.

ABNORMAL ACCUMULATIONS OF FORAMINIFERAL TESTS

Kullenberg (1952, p.26) and Parker (1958, p. 277) wondered about the abnormal accumulation of *Globigerina eggeri (=Globoquadrina dutertrei)* and both stated that these "floods" bear no relation to climatic conditions. These accumulations occurred in connection with deposits from euxinic phases and were thus caused by factors not directly related with climatic change. This abnormal fauna is characterized by (i) a large number of tests, (ii) high frequency of *Globoquadrina dutertrei*, and (iii) a low frequency of *Globigerinoides ruber, Neogloboquadrina (Globigerina) pachyderma,* and *Globorotalia truncatulinoides.* Further, *´Globigerinoides ruber´* consisted exclusively of *G. ruber ruber* during the stagnant phases, while the subspecies *G. ruber gomitulus* had disappeared during the stagnant phases. When the latter returned after an euxinic phase its size was first smaller than before (Olausson 1965, p. 231). Since *Globigerinoides ruber* (d´Orbigny) and *Neogloboquadrina pachyderma* lives in near-surface waters during the juvenile stage (Bé and Hamlin 1967) while *Globoquadrina dutertrei* is supposed to live at lower levels, the foraminifers with a moderate depth habitat were apparently favoured by the stagnations, while those with shallow habitats are retarded by the low(er) surface salinity. As stated above the ∂18-analyses suggested that the low-salinity layer was mainly confined to the uppermost ca. 50 m of the sea water of the EM.

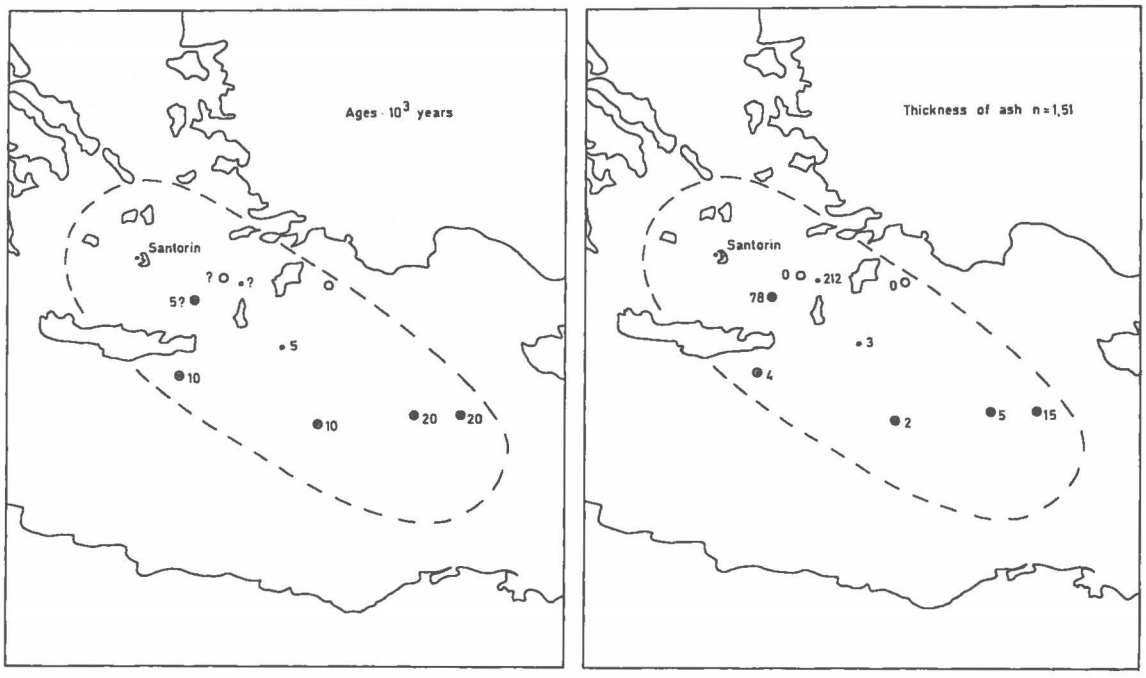
A further factor which could have promoted increased production of *G. dutertrei* is the increased organic production which has occurred during the stagnant phases (cfr. Berger, 1976). The present EM water belongs to the most oligotrofic waters of the oceans as is the case with all basins with a negative water balance. Diatoms occur in EM sediments only in connection with sapropels (Olausson 1961 a, p. 368). The organic production in the present Black Sea is, compared to the EM, higher due to the larger river water discharge and to the large amount of nutrients recycled back into the photic zone from surface waters. This supply of more nutritious water from the BS to the EM reached a maximum during the stagnant phases of the EM. This up to five times larger contribution of Black Sea water during the euxinic phases of the EM raised the organic production in the EM, producing more diatoms but retarded the growth of foraminifers in the less saline surface water. This in turn, stimulated the production of *G. dutertrei* in the more saline subsurface water layer.

Diatom mud is rare in the eastern Mediterranean. However, a layer of diatom mud is found in Core 192, formed during the strong euxinic phase S-5. It contained, according to Kolbe (priv. com.) 49 planktic species and 40 benthic and epiphytic species. A few fresh- and brackwater species as *Achnantes brevipes, Actinocyclus ehrenbergii* were found as well.

Volcanic ash beds. Mellis (1954) discussed two layers of volcanic ash in the eastern Mediterranean cores. He distinguished an upper layer of vitric ashes with a refractive index of 1.51, and a lower layer with n = 1.52. According to Mellis the upper layer originated from the eruption of Santorini some time between 1800 and 1500 B.C. Later Parker (1958) and Olausson (1961 a, 1971 b) showed that the upper layer (n = 1.51) in Core 187 and 188 was much older than the upper ashes in Cores 189 and 192. The afore-

mentioned ash beds were found to be about 20,000 years old (Olausson 1961 a, p. 372). The lower ash horizon (n = 1.52) was deposited some 25,000 B.P. (Fig. 69). Ash layers occur also at about 55,000 B.P., in the isotopic stage 6 (the Warthe substage or ≈150,000 B.P.) and in a somewhat older layer (> 150,000 B.P.).

A new approach to the study of ashes in the eastern Mediterranean cores was made by Ninkovich and Heezen (1965). They studied some Lamont cores from the Aegean Sea and the Levant and compared them with the results obtained by me. They constructed a distribution chart of the Santorini eruption in 1,500 B.C. and discussed the influence of the eruption in a very uncritical way. All ashes assumed by these authors to be from the Santorini eruption in Late



71. Left: The ages of the vitric ash with n = 1.51. The Santorin eruption occurred ca. 1625 B.C. Consequently, the ash found in deep-sea cores is much older than the ash produced in Minoan time. – Right: Thickness of ash beds shown in centimetres. The different thicknesses may depend on different distances to the sources of the ash and/or to postdepositional mixings and transport.

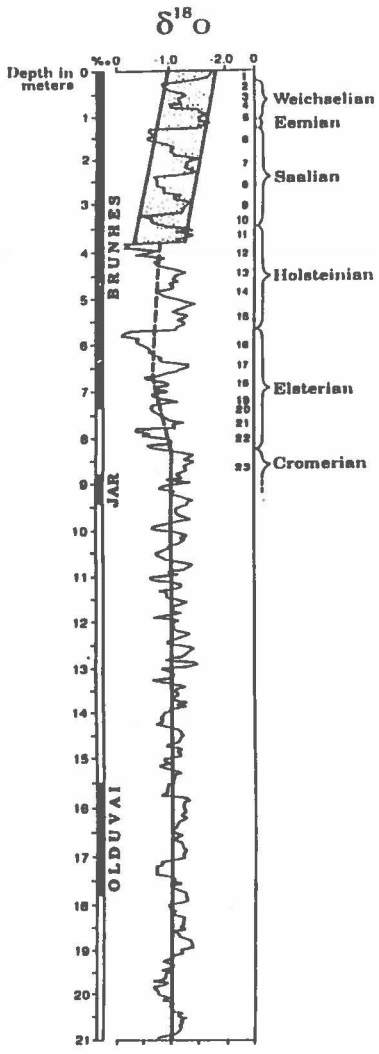
Minoan time are older than 3,500 years, and moreover, they originate from three different eruptions: from 4,000–5,000 BP, from ≈ 10,000 B.P., and from ≈ 20,000 B.P. (Fig. 71). Nonetheless some scientists have believed their map to be a serious study, having been reproduced in some textbooks as that by e.g. Turekian, (1968).

The age of the latest Santorini volcanic eruption (Bronze Age) is considered to be about 250 years older than the final destruction of the palace of Minos at Knossos, Crete, (1,375 B.C.) i.e. about 1,625 B.C. The ash did not cover Crete, as suggested by the supporters of the "Atlantis" legend. Only the easternmost part of Crete got a few mm of ash but the effect of the Bronze Age eruption, except at Thera, was very limited (Fig. 70). The discussions about tsunamis etc in connection with the Santorini eruption (Ninkovich & Heezen, op cit.) are pure speculations.

A Late Pleistocene rise in sea-level?

As stated above, the eastern Mediterranean stagnations occurred in connection with the last three interglacials (Holsteinian, Eemian, and Holocene) and during some major interstadials. During each of these periods an inflow of sea water into the Black Sea occurred, in contrast to the conditions during the Cromerian and older Pleistocene interglacials. A drop in the sill depth and/or an erosion in the channel between the Black Sea and the Mediterranean could explain the late Pleistocene EM-BS connection. However, another more likely explanation will be offered in the following.

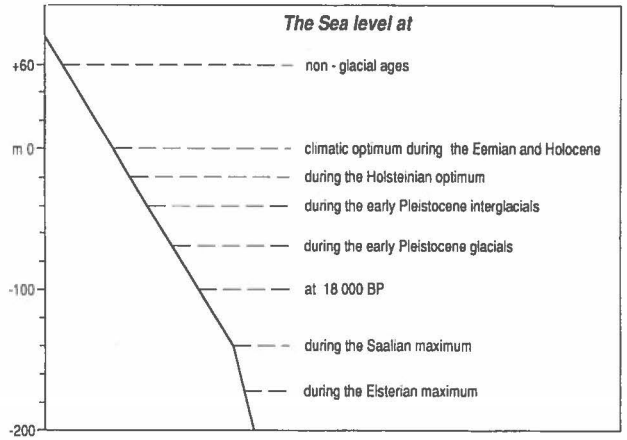
The early Pleistocene up to the end of the Cromerian interglacial (isotopic stage 23/22, alternative 21/20), was characterized by several, less intense



72. The oxygen-isotopic and paleomagnetic records in core V28-239 from a depth of 3490 m in the equatorial Pacific (3°15´N, 159°11´E) according to Shackleton and Opdyke (1976). The correlation with the European Pleistocene stratigraphy is according to Olausson (1982, 1991). Note that the $\partial^{18}0$ fluctuates around –1 per mil from the late Pliocene to the end of the Cromerian interglacial (isotopic stage 23). Then the minima reach higher $\partial^{18}0$ values. From the late Holsteinian (isotopic stage 12) there is a more or less continuous decline of the $\partial^{18}0$ upcore to the Holocene age, which is attributed to a reduction of the Antarctic ice sheet of about 15 x 10^6 km^3 and a sea level rise of about 40 m. From Olausson, (1991 b).

glaciations with intervening interglacials. The extent of the ice sheets in the northern hemisphere during these early Pleistocene glacial ages is not known. Judging by the $\partial^{18}0_{calcite}$-values (Fig. 72) and the absence of their margins in the form of end moraines etc, these ice sheets must have much smaller than those from the subsequent three glacial ages (the Elsterian, Saalian and Weichselian; see maps 83, 85 and 87). It appears from the $\partial^{18}0_c$ (Fig. 72) that the $\partial^{18}0$ in foraminifers fluctuates with an amplitude of 0.6 o/oo around the mean value of –1 o/oo(±0.3) from the late Pliocene to the late Cromerian (isotope stage 23). The isotopic amplitude then doubles from 0.6 to ca 1.2 o/oo. From isotopic stage 12 up to the Holocene the maxima and minima of the $\partial^{18}0_c$ decrease by 0.5–0.7 o/oo. A decrease by 0.5 o/oo in planktic foraminifers means a reduction of the $\partial^{18}0$ of the ocean water by this amount during these 200,000–300,000 years. A temperature rise by three degrees during this time period characterized by severe glaciations is not probable. Whence did this water come?

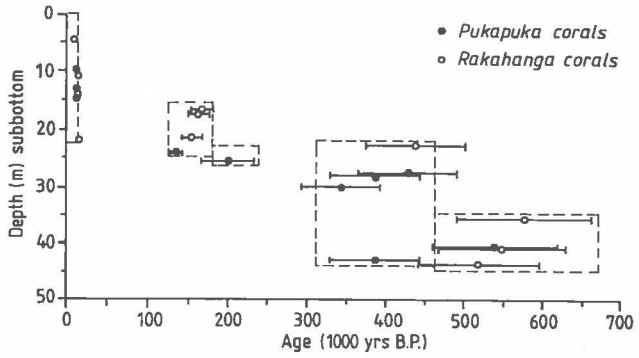
According to Robin (1987) the ice volume on the Antarctic continent was some 40 x 10^6 km^3 in late Miocene epoch. Since then the volume has diminished by about 10–15 x 10^6 km^3 due to the Pleistocene global cooling. The reduction of ice is *inter alia* due to the fact that the snow accumulation rate on the Antarctic continent is much lower during cool than during warm ages (e.g. Robin, 1987). If we assume a melting of ice in the Antarctic by about 15 x 10^6 km^3 with a $\partial^{18}0$ value therein of – 40 o/oo we arrive at the aforementioned change of $\partial^{18}0$ of the sea water (0.5 o/oo). Furthermore, a supply of 15 x 10^6 km^3 of water to the oceans is equal to a 40-m rise in sea level. This is the sill depth at the present Bosporus (and nearly that of the Bering Strait as well). Thus, the sea level stood at approximately –40 m during the Cromerian interglacial, and reached its lowest level during the intense Elsterian glacial age, being at higher levels during later glacials and interglacials (Fig. 73). A similar rise of sea level can also be deduced from Fig. 74 if we reinterprete the data by Gray et al. (1992).



73. Pleistocene sea level stands. From Olausson, (1991 b).

High sea level stands during the Eemian have been documented (see further, e.g. Flint 1971, pp.

329–343). This does not necessarily apply for the Holsteinian, which may suggest that the aforementioned melting of the Antarctic ice ended during the Saalian ice age. Since cool climate prevailed from the Elsterian to the end of the Saalian more or less continuous reduction of the Antarctic ice sheet during most of that time may have occurred..



74. Age versus depth of coral samples from Pukapuka (circles) and Rakahanga (squares). Ages less than 300,000 yrs B.P. are based on U-series analyses, and ages greater than 300,000 yrs B.P, are based on ERS analyses. Note the sea level rise by ca. 40 m during the post-Elsterian times. Data from Gray et al., (1992).

The conclusion is therefore that the very intense cooling and glaciation in the north during the late Pleistocene epoch resulted in a partial melt of the Antarctic ice, due to the lower snow accumulation rate. This caused a rise in the sea level to such an extent that the Black Sea – Eastern Mediterranean connection was established. Thus, water exchange events triggering anoxic phases of the eastern Mediterranean as discussed above, were the result of this Antarctic melting and associated global sea level rise. This sea level rise will be treated further in chapter XIII in connection with an attempt to explain the cause of the last three intense glacial ages.

The evidences from the western Mediterranean cores

There are two important studies about the western Mediterraean cores 201–211: one performed by Professor Erik Norin (1958) which focused on the volcanic sediments in the **Tyrrhenias Sea** and another by K. Gösta Eriksson (1965) dealing mainly with cores from **the Balearic and Alboran Seas** in which eolian sediments and turbidites were common. A summary of the composition of the cores 201–211 is also given by myself (Olausson 1960 c).

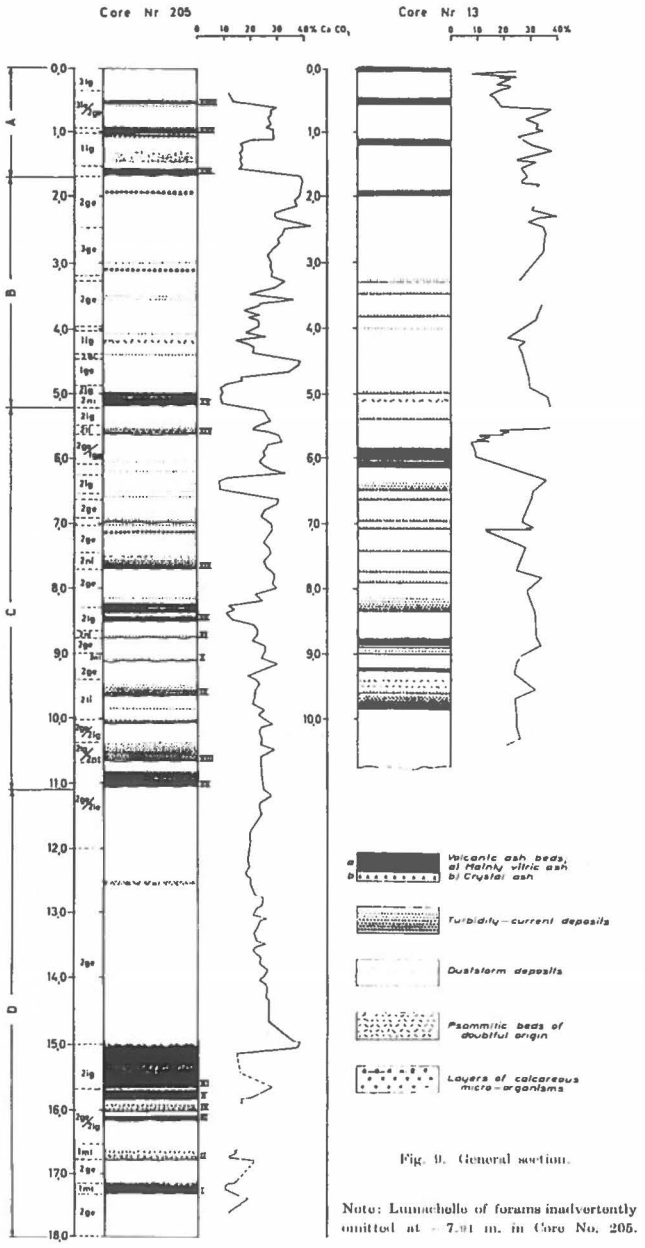
Erik Norin, formerly an expedition leader of the Sven Hedin Expedition to Central Asia 1927–1935, was chiefly a field geologist. His main interest was sedimentary petrography. In 1957 Norin received the Björk's prize (Björkénska priset) from the University of Uppsala for his studies of the sediments from the Tyrrhenian Sea.

In 1951 Norin took over the studies of Core 205 which had not yet been completed by Dr Otto Mellis, Stockholm. The ideal of complete sequences with successive layers of older volcanic ashes, intercalated in normal pelagic sediments had already been disproved by the cores from the Skagerrak Expedition in 1946: sediment sliding, turbidity currents, burrowing organisms had destroyed such an ideal stratigraphy both in the Indonesian (Olausson 1060 b) and in the Tyrhennian Sea. Only in the Eastern Mediterranean have discrete volcanic beds been found. Fig. 75 gives the stratigraphic interpretation of the two cores studied in detail.

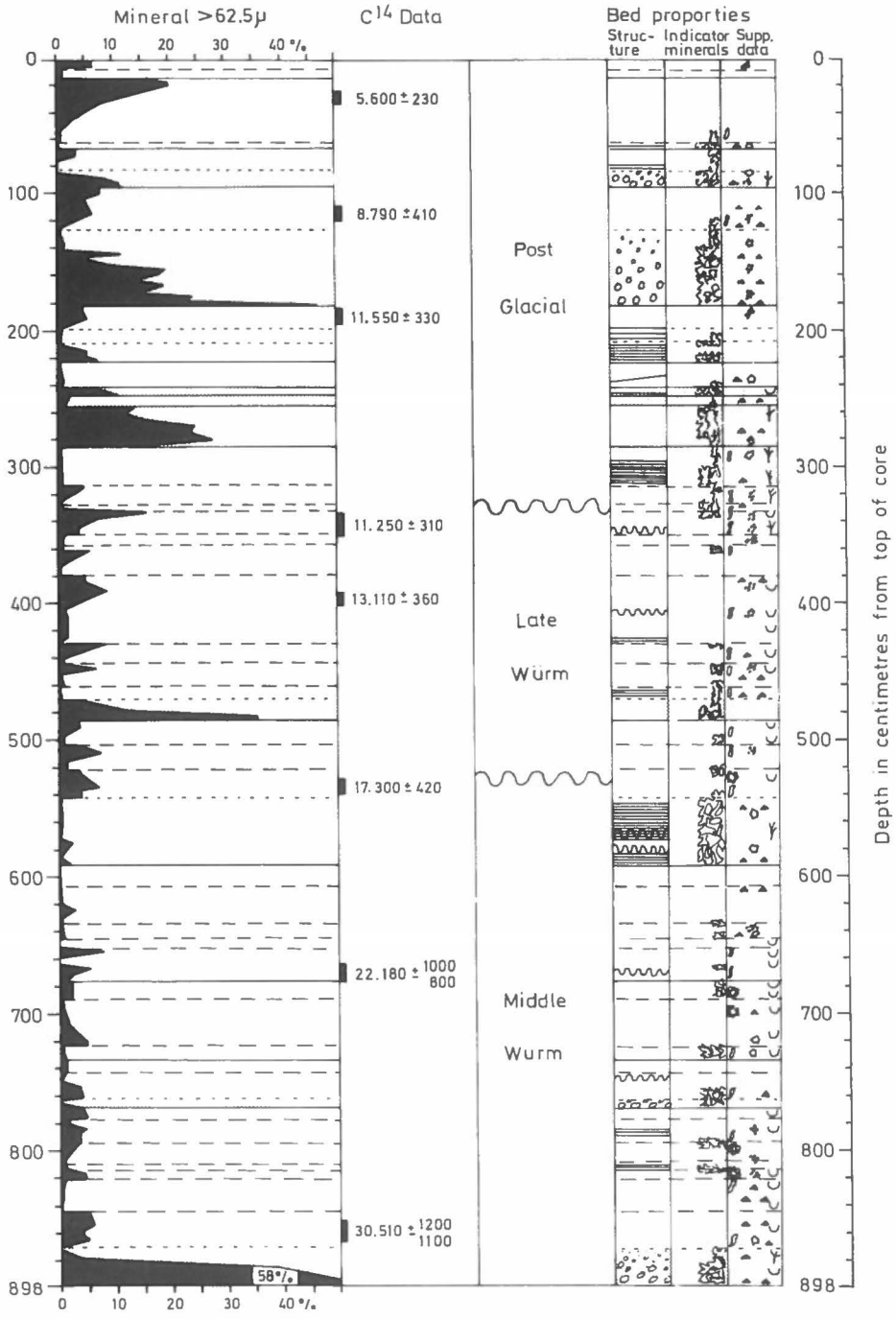
In Core 205 two components of different origin could be distinguished, viz. 1) "normal" pelitic sediments and 2) episodically deposited psammitic material. The pelitic fraction consisted of clay and fine silt particles which had reached the marine environment by river waters or by wind. Besides, coccolithophorids and some foraminifers and fragments from these. The psammitic beds intercalated in the "normal" sedimentary sequence have a thickness ranging from a few millimeters to several decimeters. These coarse-grained beds have two different origins. One type consists almost exclusively of volcanic ash from major volcanic eruptions. The other type consists of turbidites of terrigenous detritus and foraminifers. The ashes originate from the Campanian volcanic district. The well-known Plinian eruption of the Young-Somma (Vesuvius) A.D. 79 have left only thin traces in the sedimentary sequence of the distal region of Tyrrhenias Sea from which Core 205 originates. This may be due to the northwesterly wind having prevailed during the eruption which carried and deposited the lapilli mainly over the south-eastern quadrant of the Somma (Norin, op.cit. 99). Aforementioned eruption may have formed the ash bed no. XV at a core depth of 5 m in Core 205 (Fig. 75). The entire core cover the Weichselian – Holocene time span.

The sediments from the **Balearic – Alboran Seas** are different from those in the eastern Mediterranean. As referred to before a layers of sapropel layer of S-5 age is found in Core 201 which is the only core in the western Mediterranean extending further back in time than the last ice age.

As appears from Fig. 76 (redrawn from Eriksson 1965) the core 210 (as 209 and 211) represents a short period of the Quaternary period: only the last approx. 100 000 years. According to Fig. 76 one third or more of the core material could be of terrigenous origin. This component has been transported to the sea either by winds as indicated by the microstratified wind-born sediments in the Weichselian – early Holocene part of the core, or by turbidity currents. Five turbidites are marked on Fig. 76. "Desert quarts" is the dominating eolian component. The clay minerals are from two main provinces: quasi-illite and kaolinite are of Saharan origin, and illite and chlorite are of northern origin. The pelagic component is represented by calcareous matter, constituting 20–75 % of the bulk sediment. There is no distinct difference between the sediments from the last ice age compared with the Holocene, except for foraminifers. *Neogloboquadrina pachyderma* is frequent in the glacial stage and *Globorotalia inflata* and *Globoquadrina dutertrei* increase in frequency in the Holocene stage.


Core Nr 205
Core Nr 13
Volcanic ash beds,
a) Mainly vitric ash
b) Crystal ash
Turbidity-current deposits
Duststorm deposits
Psammitic beds of doubtful origin
Layers of calcareous micro-organisms
Fig. 9. General section.
Note: Lunulichelle of forams inadvertently omitted at – 7.91 m. in Core No. 205.

75. The stratigraphy of Cores 205 and 13 from the Tyrrhenian Sea. From Norin, (1958).



76. The stratigraphy and lithology of the core 210 from the Balearic Sea (western Mediterranean). From Eriksson, (1965). Note the six turbidites and ten microstratified beds of eolian matters (loess).

XIII. On the cause of the last three intense glacial ages

On the general cause of Cenozoic climatic changes

Before discussing the cause of the late Cenozoic climatic deteroriation I will make two general remarks:

a). In this attempt to explain the cause of the last three intense glacial ages I will **exclude** – as I did before – the possibility that changes in the orbit of the Earth are the cause of the glacial – interglacial cycles. The Milankovitch hypothesis has never been proved. The recent discovery of the 100,000-year cycle in the accretion rate of cosmic dust to the Earth, refered to in chapter VII, is a warning to take the Milankovitch hypothesis for granted. What extraterrestrial factor, which may control long-term climatic variations, is left open here. Moreover, I point out once more, that the present Plio-Pleistocene chronostratigraphy, based on the Milankovitch hypothesis, must be considered as only approximate.

b) In Chapter VI:3 C it is stated that the controlling factor of deep water formation, the fresh water supply, to the Nordic Seas was about one-fourth of the present amount during ice ages. This decrease together with the increased cooling, must have favoured downwelling there. Furthermore, high pressure prevailing over Greenland existed since the Saalian ice age (stages 10-6) has been responsible for concentrating the fresh water input to the peripheral part of the Nordic Seas. This favours the central part to be open (Stigebrandt, 1985, p. 302). It is difficult to understand from physical point of view how the vertical circulation of the North Atlantic had stopped during such extreme climatic conditions. – It is also difficult to understand how the flux of the Antarctic Bottom Water could have increased in the Atlantic during the ice ages while it remained nearly constant into the Indo-Pacific Ocean – I do not understand how an increased equatorial upwelling and a doubled carbonate accumulation could occur in the Indo-Pacific Ocean during ice ages without an increased downwelling and dissolution in the North Atlantic. – The Cd/Ca-data of benthic foraminifers can be interpreted in the two (opposite) ways, i.e. also in favour of my theory – I am not able to see that e.g. the $\partial^{13}C$- data, presented by Raymo et al, 1990, can be interpreted in the way as done by the authors. Their $\partial^{13}C$-analyses indicate a continuous deep water formation in the Nordic Seas throughout late Cenozoic times except for intense deglaciation phases when it ceased (see Fig. 30 C).

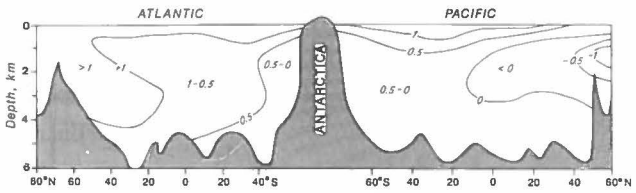
The cause of the long term climatic cycles

The cause of the long-term climatic cycles as the glacial – interglacial cyclicity is, as stated above, extraterrestrial and this fluctuating heat source is considered as the **primary factor** in the climatic development.

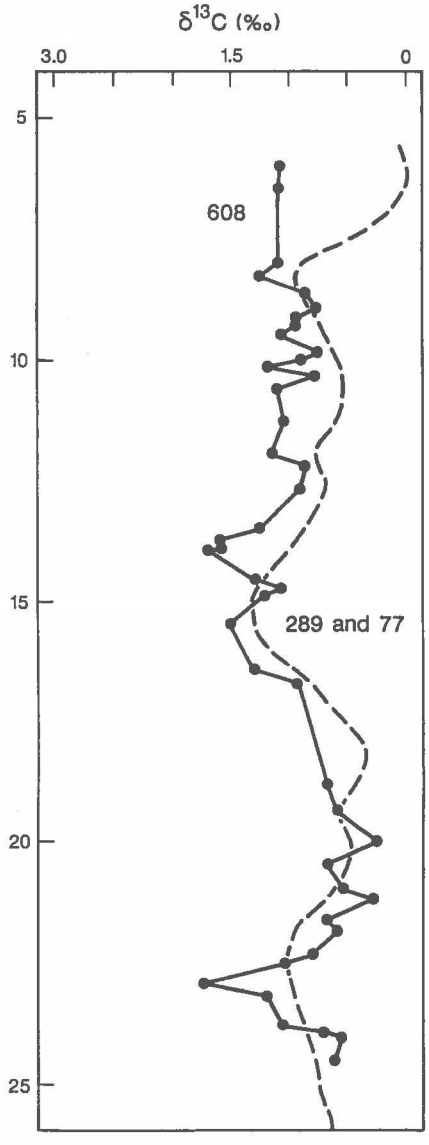
Then, there are **secondary (terrestrial) factors** which strengthen or weaken the cosmic influences. Most important secondary factors in this connection are the geographical distribution of continent(s) at high-latitude and the presence of salinity stratified water bodies of the polar – subpolar regions.

It is well-known that the thermal isolation of the Antarctic continents ca 35 Ma ago, after the separation of this continent from Australia, led to heavy glaciation of eastern Antarctic and to an intensified bottom water formation near to Antarctica. A further isolation by opening of the Drake Passage about 21 Ma ago and the establishment the West Wind Drift causing further ice growth. The Antarctic ice sheet reached a maximum, ca 40 x 10^6 km^3, in late Miocene times (Robin 1988). The size of the ice sheet from then on then depending on the snow accumulation rate which is larger during warmer times, less during cool ages (Robin 1988, Fig. 2–3).

High (positive) $\partial^{13}C$-values of benthic foraminifers are in the present ocean the characteristic of young deep water of the "Atlantic type" (see Fig. 77), while the 1–1.5 o/oo lower, mostly negative, ∂-values indicate old bottom waters of the "Pacific type" (see further the discussions by e. g. Olausson, 1988, p.112–114). The pattern of the ocean circulations

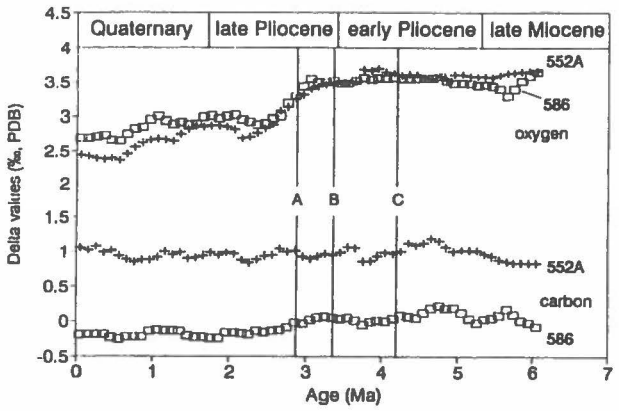


77. The $\partial^{13}C$ in the ocean water from the North Atlantic to the Pacific. Data from Kroopnick, (1985). Redrawn from Olausson, (1988, Fig. 9).



78. Age comparison of isotopic records of benthic foraminifers from the Pacific Sites 289 and 77 (dashed line) with the eastern North Atlantic 608 (simplified from Miller et al., 1987, Fig. 4).

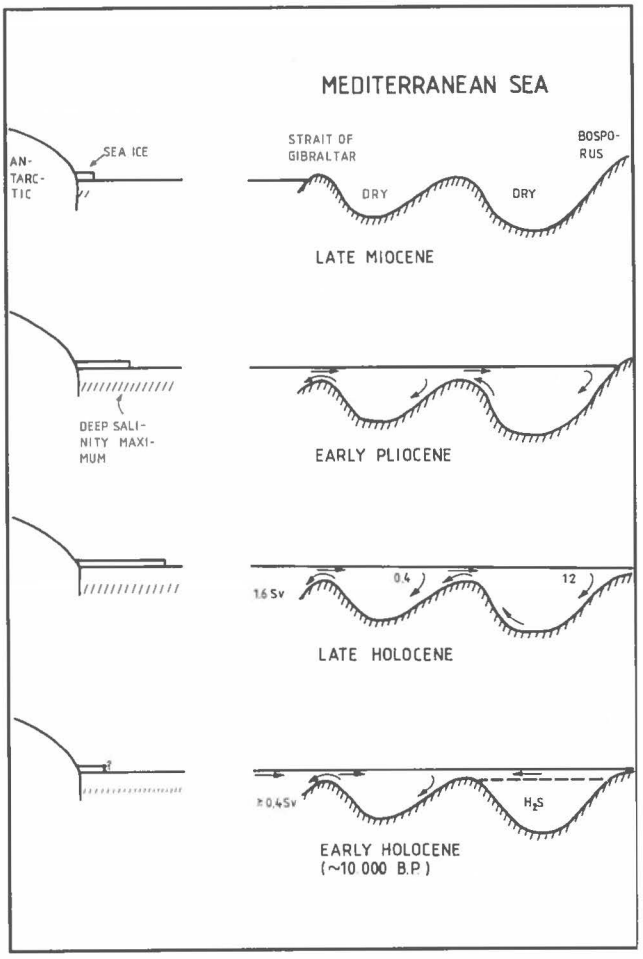
remained rather unchanged during the Oligocene – Miocene. At about 6.2 Ma the deep water circulation changed. Equatorial Indo-Pacific cores (e.g. RC12-6, Sites 289, 677; Woodruff et al. 1981, Vincent et al. 1985, Raymo et al. 1990) began to have more $\partial^{13}C$ change from 0 to –1 o/oo of the benthic foraminifers, while those from the Atlantic (e.g. sites 552 A and 608 Shackleton and Hall 1984; Miller et al., 1987; Raymo et al., 1990), at the same time began to have more positive $\partial^{13}C$-values. The bottom water formation in the northern Atlantic Ocean began thus ca 6.2 Ma ago (Fig. 78). This trend is also obvious in the study by Whitman and Berger (1992). They showed (op.cit., fig 10–11) that the $\partial^{13}C$ values of the benthic foraminifer *Planulina wuellerstorfi* were always at least 1 per mil heavier (and positive) in the North Atlantic Site 552 A than in the Pacific core 586 from the Ontong-Java Plateau with a tendency of a larger difference after the major cooling at 3.2 Ma. Core 586 shows only negative $\partial^{13}C$ values after that cooling (Fig. 79). The $\partial^{13}C$-values in Site 586 exhibit that the North Atlantic has been a permanent source of descending water masses through the whole Plio-Pleistocene epochs with a larger $\partial^{13}C$ after 3.2 Ma when the carbonate transport from the the North Atlantic began. – There is a tendency that the $\partial^{13}C$-values has been somewhat more positive in both Atlantic and Pacific cores during the last half a million years or so (e.g. Raymo et al. op. cit. fig. 2), a trend which suggests higher rate of bottom water fluxes than during earlier times. This agrees well with the carbonate and diatom data discussed above. – The cause of the circulation change at 6.2 Ma is generally explained as due to the lowering of the sills between Greenland and Scotland favouring water exchange between the Nordic Seas and the North Atlantic.



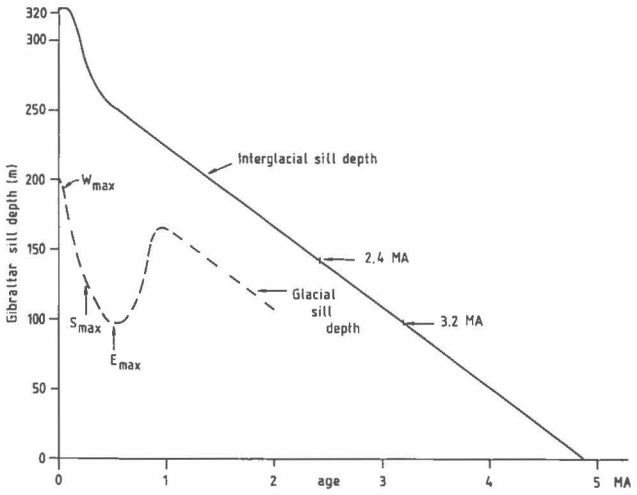
79. Comparison of the North Atlantic isotope record (Site 552) with Ontong Java record (Site 586). From Whitman and Berger, (1993, Fig. 11).

THE MEDITERRANEAN AND CLIMATIC CHANGE

The ancient Tethys Sea had a deep water connection with the Atlantic Ocean up to ca. 8 Ma ago and deep-sea carbonates had accumulated in the Mediterranean basins. Between 8 to 6.14 Ma (i.e. late Tortonian) the circulation became restricted (see e.g. Thunell et al., 1987). The following Messinian salinity crisis, 6.14–5.26 Ma (Hsü et al., 1977; Aharon et al. 1993) was, as far as I understand, caused by a hot spot below the Gibraltar area during that time. When the area (plate) moved away from that hot spot, which presently is located below the volcanic islands outside the NW Africa, the sill sank below the sea-level of the Zanclian age. When the water influx from the Atlantic was larger than the evaporated water mass within the Mediterranean basin (> 0.16 Sv) the injection of high salinity water into the Atlantic Ocean began (Fig. 80). This injection favoured the formation of the deep water in the Nordic Seas and resulted gradually in the formation of the deep salinity maximum between ca 25°N and 70°S in the Atlantic Ocean. If we assume a continuous subsidence of the Gibraltar sill we can illustrate this downward movements as at Fig. 81. The rate of subsidence could have been in the order of 50 m per 10^6 yrs. If this is correct correct the sill depth at Gibraltar at ca 2.4 Ma could have been as at 18,000 B.P. However, the salt export could have been higher at 2.4 Ma because of higher evaporation and thus higher salinity of the outflowing water.



80. The pack ice surrounding the Antarctic results from the salinity stratification in the underlying waters. The deep salinity maximum has its origin in the effluent, very dense, Mediterranean water. During the late Miocene (the Messinian "Salinity crisis") this efflux was stopped, and the during the early Holocene stagnation in the eastern Mediterranean the outflow diminished considerably. The early warming in the Australia-New Zealand area (Salinger), preceeding the warming of the northern hemisphere by about 3,000 years, and the abnormally early isotope maximum in the Antarctic ice dome C (Lorius et al., 1979) may have been caused by this reduced outflow from the Mediterranean. This reduced outflow started the following reaction in the southern hemisphere: weakened salinity stratification -> reduced extent of sea ice around the Antarctic -> lower albedo of the Southern Ocean -> climatic optimum in the Southern Hemisphere.1 Sv (Sverdrup) = $10^6 m^3 sec^{-1}$. From Olausson, (1982).



81. Theoretical water depth of the Gibraltar Strait during the last 5 million years. The salinity effect on the North Atlantic water could have been higher than shown only by the water depth at the sill. The evaporation could have been twice as large during the warm Pliocene (up to 2 m/yr) than today (c. 1 m/yr). Thus the salt export could than has been c. 90 x 10^9 ton larger per year in early post-Messinian period than today. – E_{max}, S_{max} and W_{max} stand for maximum of the three last glaciations, Elsterian, Saalian and Weichselian respectively.

As stated in chapter VI:3 D there was an increase in the deep water formation of the North Atlantic during the late Pliocene, at first at approx. 3.2 and more pronounced at 2.4 Ma (see Fig. 35–37). According to

Hodell et al., (1985) there was a reorganization of the deep-vertical water mass structure in the western South Atlantic at ca. 3.2 Ma (North Atlantic Deep Water increased). As stated above enhanced ocean convection is favoured by the influx of Mediterranean water which contributes to the high density required to form sink water in the Nordic Seas. It is probable that the aforementioned deepening of the Gibraltar Strait already at 2.4 Ma reached such a magnitude (Fig. 81) that the outflow of high-saline water began to increase the North Atlantic convection.

On the Bering Strait and its water exchange

The present fresh water cycle is illustrated in Fig 34. Fresh water (vapor) is transported in the air from the North Atlantic to the Pacific, particularly at low altitudes and during the northern winter (Weyl, 1968), thus increasing the surface salinity in the North Atlantic and reducing it in the North Pacific. This transport facilitates the deep water formation in the North Atlantic and is one of the factors preventing it in the North Pacific. The fresh water carried to the Pacific is returned to the North Atlantic via the Bering Strait-Arctic Ocean. The size of this fresh water transport is 0.1 Sv. Some 0.12 Sv river water is also discharged to the Arctic Ocean. All this fresh water creates the salinity stratification in the Arctic Ocean, the prerequisite for a permanent pack-ice cover.

It appears therefore that (1) the opening of the Bering Strait must have improved the conditions in the Arctic Ocean for sea ice formation while its closure had a negative effect on the extent of sea ice, and (2) due to the large albedo difference between an ice-covered (60–90 % ,depending on the the extent open water, i.e. polynyas) and an ice-free ocean (10–20%) this hydrographic change must have had a cooling effect on the areas surrounding the Arctic Ocean, thus initiating glaciations in its vicinity.

The key problem in order to solve the riddle of the late Cenozoic climatic deteroriation is thus the depth and the exchange in the Bering Strait. When did it open? When began the water exchange to become climatically important? If the initial surface current was directed to the North Pacific when became it reversed? When did the Polar Sea get a perennial sea ice during interglacials?

Most geoscientists studying the past water exchange between the Pacific and the Polar Sea and the migration of organisms between these oceans agree that the Bering Strait had been opened either in late Pliocene or during early Pleistocene (ca. 4–2 Ma). There is no doubt the strait was open in the Middle Pleistocene (early Brunhes).

The cause of the climate change ca 0.8 Ma

In order to look for the cause of the middle-late Pleistocene climatic deteroriation we must look for a factor which lowers the temperature globally or at least in the both polar regions, and, in addition increases the water depth in the Bering Strait. I will not here discuss the hypothesis of the plateau uplift in Southern Asia and the American West. This was to be the main cause for the late Cenozoic northern hemisphere cooling as suggested by Ruddiman and Kutzbach, (1989). This uplift has certainly a real effect but it works too slowly. I need in this connection something which works faster and, besides, causes a rise in the sea level as well.

The sea ice around the Antarctic is a result of the salinity stratification around the continent (Fig. 80). This is caused by the deep salinity maximum, which is the most important factor, and by precipitation, and melting of ice bergs. According to Weyl (1968) the sea ice formed around the Antarctic has a theoretical thickness limited to 1 m. The deep salinity maximum originates from the Mediterranean outflow through the Strait of Gibraltar. This water spreads in the North Atlantic at a water depth of about 500 m, passing the equator at about 1,500 m depth and reaches the Antarctic region at a water depth of about 500 m. The albedo of an open Southern Ocean is about 20, while an ice-covered surface reflects 60–90 % of the received radiation. I have earlier suggested that the Pliocene cooling was a results of the development of the deep salinity maximum in post-Messinian times (Olausson 1982, p.12, redrawn in Fig. 80). Judging by Fig. 81 it seems likely that the outflow of high-saline water could have become of importance already c. 2.4 Ma and could have been enhanced near the Matuyama/Brunhes transition.

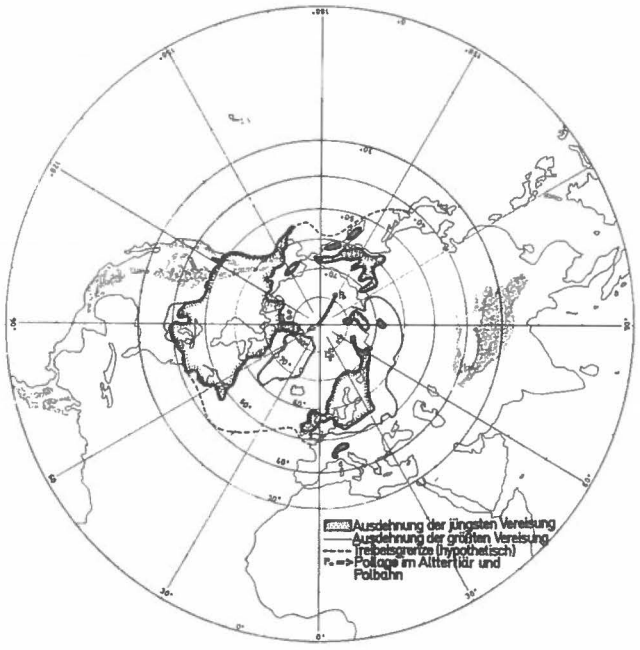
Some time around 1.2 Ma the amplitude of the $\partial^{18}0_c$ in Central Pacific cores (e.g. V28–239, Shackleton and Opdyke, 1976; Site 586, Whitman and Berger, 1992; Site 805, Berger et al., 1993) increases and the amplitude increased considerably upcore from the isotopic stage 22, i.e. from the

Elsterian glacial ice onwards (Fig. 72 and 79). From the isotopic stage 12 (late Holsteinian) there is a more or less continuous decline of the $\partial^{18}0$ towards the Holocene, which could be attributed to a reduction of the Antarctic ice sheet of about 15×10^6 km^3 and a sea level rise of about 40 m (Olausson 1991 b). It is my conclusion that the development of the deep salinity maximum created a larger sea ice area around the Antarctica, and, due to increased albedo of the Southern Ocean, led to a gradual cooling, at first in the southern hemisphere, later on globally. During the long time interval from the Elsterian to the end of the Saalian (= isotopic stages 20–5 e; three–fourth of this period was glacial) the accumulation of ice on the Antarctic was reduced.

The lower rate of snow accumulation resulted in a partial melt of the Antarctic ice sheet (see Robin, 1988). Of the former (Miocene) ice sheet remained ca. two-third and the melting caused a sea level rise by ca 40 m (see further Chapter XII). This excess melting has also temporarily contributed to sea ice formation in the Southern Ocean.

The sea level rise caused the inflow to the Black Sea as outlined before (Fig. 67 and 73), higher sea level stands recorded on coral islands (Fig. 74), transgression into the North Sea and the opening of the English channel in the Holsteinian etc. Moreover, it must have caused a depening of the Bering Strait. The rise in sea level thus increased the transport of low-salinity water from the Pacific to the Arctic Ocean, thus strengthening the halocline there. The gradual change from mainly winter ice in the Arctic to a perennial ice cover must have reduced the precipitation in adjacent Eurasia. This is because an open Arctic may give up to 50 % higher precipitation than an ice-covered one could give. The development of a perennial ice cover in the Arctic gradually reduced the moisture in early phases of the last three glacial ages in eastern Siberia. This may be the cause of the decrease in the extent of the ice sheet in Eurasia from the largest during the Elsterian to the much smaller and westerly concentrated Weichselian ice sheet (Fig. 82, 83, 85 and 87). The moisture from which the ice sheet in Siberia was built up came from Atlantic, Arctic and, in extreme eastern Siberia, from the Pacific sources.

The corresponding North American ice sheets, the Kansan, Illinoian and Wisconsinian were rather equal in size (Fig. 82). They were nourished by the moisture from the maritime air from the southwest, later from the south (Flint, 1971) and, thus, not depending on the Arctic-Atlantic sources. The Greenland ice sheet must have become permanent during the Saalian ice age as a result of the permanently ice-covered and thus very cold Arctic Ocean when compared with the warmer only winterly ice-covered Polar sea in pre-Saalian times. The cooling effect of an ice-covered Arctic Ocean is much larger than the effect of the Greenland ice sheets, mainly due to its larger albedo.

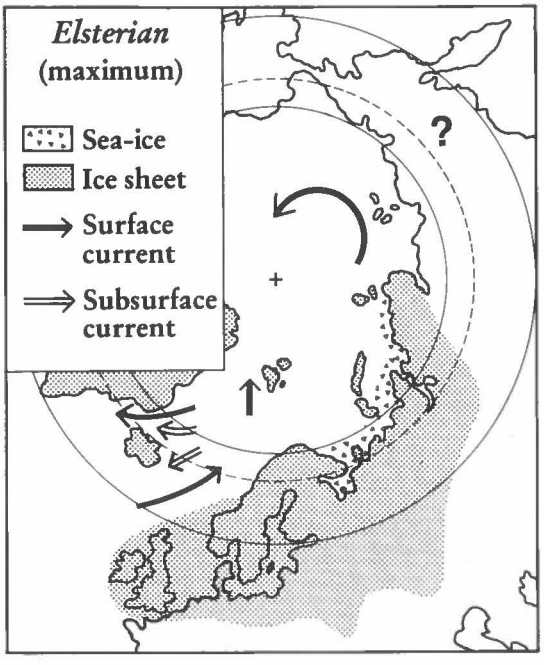


82. The extent of ice cover during the maximum of the Pleistocene glaciation (Elsterian –Kansan) and the during the Weichselian ice age. Note the difference in size over Russia and the (±) constant size over the North America. The explanation for difference is given in the text. After Flohn, (1969).

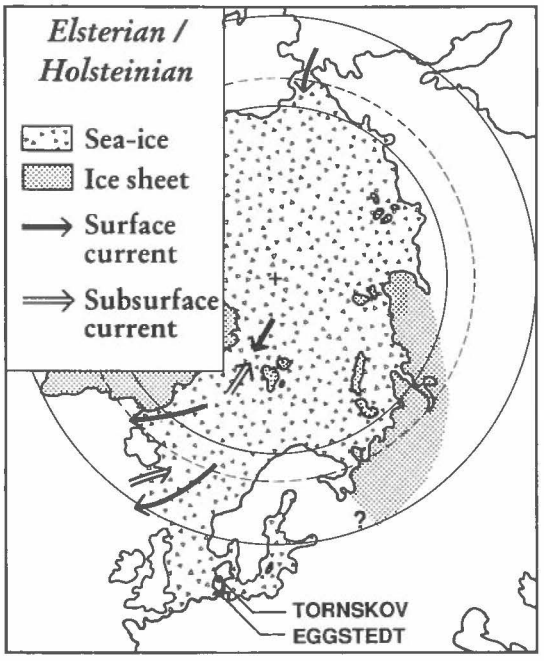
As outlined in an earlier chapter (VI:3 D b) our Core 280 from the North Atlantic suggests that the Polar Front and the zonal belts were displaced 5–7° northwards and that the Greenland ice sheet must have been smaller in pre-Saalian times. Due to the topography of Central – North America, a more northward position of the atmospheric fronts and belts could have diminished the westerly fluxes of moisture from the Atlantic to the Pacific. Thus it may have been North America instead of the central Pacific which received the evaporated water from the North Atlantic in pre-Saalian times and currents in the Bering Strait may have had the opposite directions.

After Greenland became permanently ice covered during the Saalian and the present fresh water cycle was established during the Eemian, the currents in the Bering Strait became reversed, i.e. as it is today.

Thus, the ice sheets in Eurasia decreased in size from the Elsterian to the Weichselian ages as a consequence of the increased sea ice cover in the post-Cromerian Arctic Ocean. The duration of the deglacial times after each of the these three glaciations decreased too. All three deglaciations were two-steps as they were probably interrupted due to freezing of the Arctic Ocean when the meltwater discharge and the opening of the Bering Strait reestablished the halocline there. However, while much of the ice sheets in Eurasia after the Elsterian and Saalian remained to be melted after their "Younger Dryas"-phase only a small ice-cap existed after the Younger Dryas proper (see Fig. 84, 86, and 88). Within the context it is also worth noting that the currents across the Iceland-Scotland Ridge were reversed during the post-Elsterian and post-Saalian times (as the currents in the Strait of Gibraltar during the post-Saalian stagnation S-5) but not after the Weichselian. This is due to differences in the magnitude of the meltwater discharge (see Fig. 84, 86 and 88).



83. Maximum extent of the Elsterian ice sheet in western Eurasia and the hydrography of the Nordic Seas during these times. The Arctic Ocean is supposed to have been open. After Olausson, (1995).



84. The extent of the ice of Eurasia and the hydrography of the Nordic Seas during the Elsterian/Holsteinian transition (isotopic stage 16/15). Huge amount of meltwater from the western part of the large Elsterian ice sheet as well as comparatively low sill depth at the Greenland-Scotland ridge and an (±) absence of a Greenland ice cap contributed to this spread of the meltwater masses. Low salinities of the North Sea area (Olausson, 1995) suggests a surface outflow from the Nordic Seas. This situation represents an interrupted deglaciation, a "Younger Dryas"-phase, where much ice remained to be melted (Olausson, op. cit.). The meltwater has probably caused much sea ice. After Olausson, (1995).

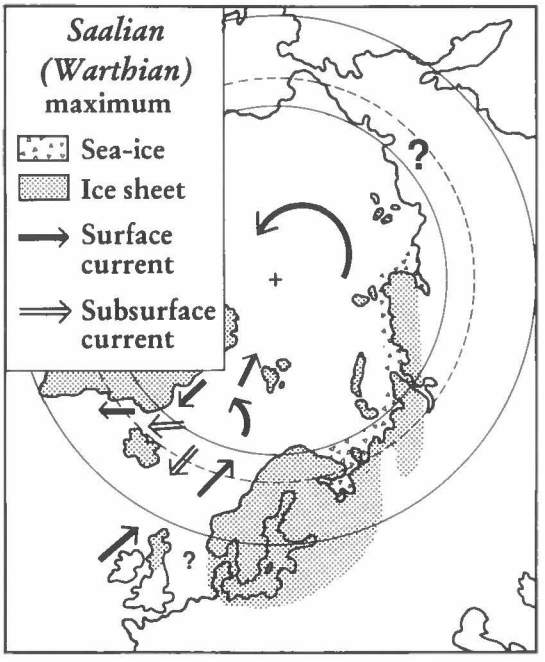


Fig. 85. The Saalian (Warthian) ice sheet in Eurasia and the hydrography of the adjacent Nordic Seas during this glacial time (isotopic stage 6). The Arctic Ocean is supposed to have been open. The extent of ice sheet in eastern Siberia is not given. From Olausson, (1995).

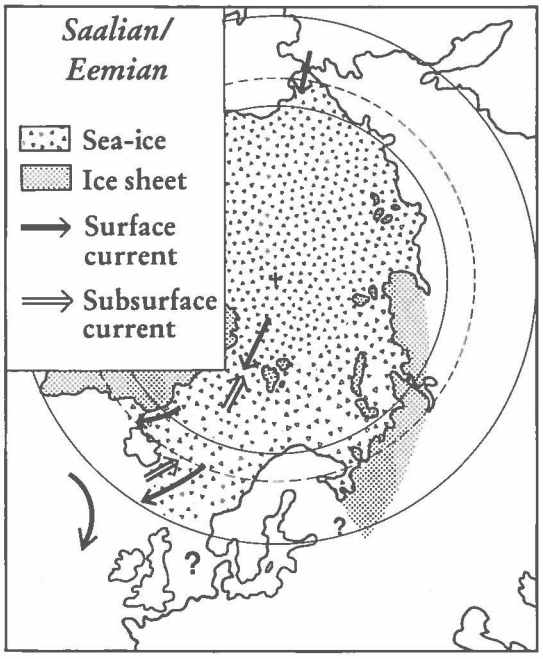


Fig. 86. The assumed extent of the remaining Saalian (Warthian) ice sheet and the hydrography of the adjacent Nordic Seas at the Saalian/Eemian transition (isotopic boundary 6/5). Evidence for low surface salinity in the Nordic Seas suggests surface outflow (Olausson, 1983, 1995).
Note the large ice sheet in the east which remained to be melted during this two-step deglacial time, a "Younger Dryas"-phase. High early Eemian sea-level is recorded in the Baltic (e.g. Forsström and Eronen, 1985).

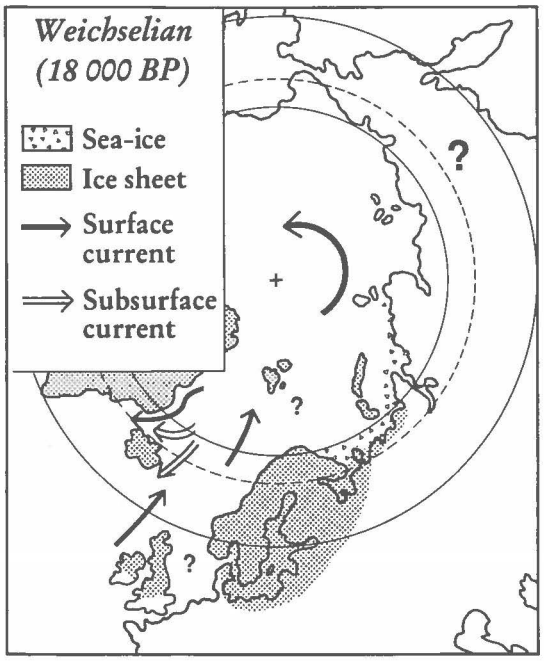


Fig. 87. Generalized map of the Eurasia and the Nordic Seas 18,000 B.P. The Arctic Ocean is supposed to have been open (Olausson and Jonasson, 1969). The ice condition in the Barent Sea and in the eastern Siberian are left open. Note that the ice sheet in the eastern Eurasia is much smaller than during the Saalian and particularly the Elsterian. This is, as suggested in the text, due to less sea ice in the Arctic Ocean during the early phase of the Elsterian than later.

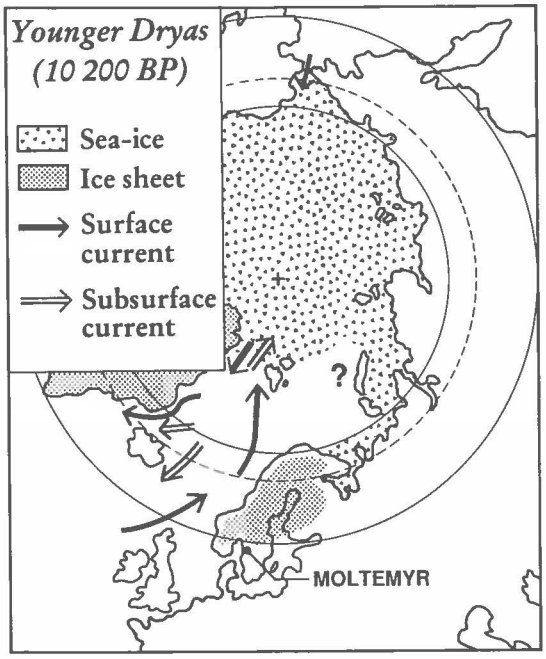


Fig. 88. Scandinavian ice sheet and the Nordic Seas at the end of the Younger Dryas time (10,200 ^{14}C B.P.). The sea ice in the Arctic Ocean is supposed to have been reestablished due to the meltwater influx during the Bölling-Alleröd interstadials and due to the opening of the Bering Strait (present sill depth ca. 40 m).

In summary:
the post-Messinian increased outflow of high-saline Mediterranean water developed gradually the Deep Salinity Maximum in the Atlantic

which at the Cromerian/Elsterian transition contributed to the formation of a large sea ice area in the Southern Ocean,

which cooled the southern hemisphere and later both high-latitude regions,

which reduced the snow accumulation on the Antarctic ice sheets,

which caused a gradual melt of the Antarctic ice sheet, such that

during the Saalian it had reduced its size by one-third of the former (Miocene) size,

which resulted in a sea-level rise by ca 40 m deepening the sill depth of e.g. Bering Strait,

which favoured the transport of less saline water from the Pacific to the Arctic Ocean,

thus strengthening the halocline in the Polar Sea and leading to the formation of thicker sea ice, and during Eemian and Holocene a perennial ice cover ,

which ice cover made a permanent Greenland ice cap possible and made the glacial Eurasian ice sheets gradually (from Elsterian to the Weichselian) smaller and concentrated more to the west ,

a permanent Greenland ice cap forced the climatic belts over the north Atlantic to a more southerly position,

so that the present fresh water cycle (Fig. 34) became established and the currents in the Bering Strait got their present directions.

It could be added that when the present inflow from the Pacific to the Arctic Ocean increased a new source of nutrients to this ocean was established. Two-third of the Si and nearly all N and P in a nutrient balance of the Polar Ocean, given by Björk (1990), originate from the Pacific. An equal amount of Si is lost to the North Atlantic. This nutrient transport and source for the North Atlantic thus must have been negligible during the Cromerian and earlier times.

THE CLIMATES DURING A GLACIAL PHASE AS CA. 25,000–7,000 BP

It has been a matter of debate whether Quaternary long-term climatic changes on the two hemispheres could have been inversely related as predicted by the Milankovitch hypothesis. For the past 0.3 Ma paleotemperature trends in the northern and southern hemisphere have not been entirely synchronous, with changes in the southern hemisphere temperatures leading those in the north by 2,000–3,000 years (Hays et al., 1976; Salinger, 1981; Jansen et al., 1986; Labeyrie et al., 1986; Kuipers, 1989). Jansen et al. (op. cit.) concluded that between 0.3 and 0.4 Ma oceanic circulation in the southern hemisphere changed from a more glacial pattern to more typically interglacial conditions. I have above given arguments how the Mediterranean has influenced the climate at the southern hemisphere which in its turn changed the climate of the northern hemisphere. I will give another example of the interrelationship between the two hemispheres.

The peak of the last glacial subage occurred ca. 22,000 BP in the southern hemisphere, while at 18,000–19,000 in the northern hemisphere (Salinger 1981). From these periods onward there was a gradual increase in temperature starting in the tropics, reaching a maximum in the Australia-New Zealand area at ca. 9,000 B.P., and ca. 6,000 B.P. in the northern hemisphere (Salinger op. cit; Olausson, 1984 Fig. 2;1985, Fig. 2). There are no exact data as to the lowest sea-level stand which presumably occurred at 18,000 –19,000 B.P. It would be safe to consider –100 ± 20 m as an approx. minimum sea-level stand. From ca. 13,000 B.P. to 9,000 B.P. (from the Bölling to the Boreal chronozone) the sea level rose steeply. Various sea-level curves indicate that at least 50 % of the northern ice sheets melted during this time-span. During this time the salinity dropped by 1–2 o/oo in the surface water of the northernmost Atlantic, creating a (quasi)stagnation there (Fig. 30 C; Olausson, 1985).

The present volume of the Mediterranean outflows is 1.6×10^6 m^3 sec^{-1}. As shown above (Chapter XII), the eastern Mediterranean was stagnant ca 11,000–7,000 B.P. The weastern basin was, so far as is known, not stagnant and the bottom-water formation and the exchange with the Atlantic (the subsurface outflow), was, thus, restricted to this basin. The western Mediterranean is, at present, responsible for only 25 % of the bottom water formation and to some 30 % of the salinity increase of the Mediterranean surface water (due to evaporation). The volume and salinity of the outflowing water could have been appreciably lower than today (Fig. 81). This could be due to transport of low-saline surface water from the eastern to the western Mediterranean. Therefore, the deep salinity maximum could have been more or less absent in the south Atlantic during a few thousand years around 9,000 years B.P. This lack of salinity stratification should in its turn, have appreciably reduced the area of the sea ice around the Antarctic and, due to changes in albedo (from about 70 to about 10 %) during southern summer, improved that climate in the southern hemisphere. The extent of the sea ice around the Antarctic is more than twice that in the Arctic Ocean and any changes would, thus, exert a stronger influence on the climate than would the Arctic sea ice. The effect of the latter is equal to a temperature change of ca. 2°C of northern hemisphere during the northern summer.

The climatic optimum occurring 10,000–8,000 years B.P. in Australia and New Zealand (Salinger, op. cit.) and the contemporary isotope maximum in the Antarctic ice Dome C site (Lorius et al., 1979) could thus reflect the eastern Mediterranean stagnation and, indirectly the deglaciation in the north.

The ultimate/primary cause of the glacial-interglacial cycles is not discussed here but it should be borne in mind that the initial cooling which led to an ice age occurred very suddenly, in less than 1,000 years (Fig. 61 and Chapter X; Olausson et al., 1971).

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XV. REFERENCES

Aldahan, A. A., G. Posssnert, and E. Olausson, 1992: ^{10}Be distribution in sediments of two cores from the North Atlantic: an approach for evaluating environmental change. Bull. Geol. Inst., Univ. Uppsala, N.S. 16, 69–80.

Aharon, P. et al. 1993: Sea-level events in the South Pacific linked with the Messinian salinity crisis. Geology, 21, 771–775.

Arrhenius, G., 1952: Sediment cores from the east Pacific. Rep. Swed. Deep Sea Exped. 1947–1948, 5: 288 pp, 3 maps.

Arrhenius, G., G. Kjellberg, and W.F. Libby, 1951: Age determination of Pacific chalk ooze by radiocarbon and titanium content. Tellus 3, 222–229.

Arrhenius, G., and E.D. Goldberg, 1955: Distribution of radioactivity in pelagic clays. Tellus 7, 226–231.

Arrhenius, G., 1988: Rate of production, dissolution and accumulation of biogenic solids in the ocean. Palaeogeogr. Palaeoclimatol. Palaeoecol., 67, 119–146.

Arrhenius, G., G. Kjellberg and W. F. Libby, 1951: Age determination of Pacific chalk ooze by radiocarbon and titanium contents. Tellus, 3: 222–229.

Bé, A. W. H., 1977: An ecological, zoogeographical and taxonomic rewiew of recent planktonic Foraminifera. *In:* A.T.S. Ramsay (ed.): Oceanic Micropaleontology, Academic Press, New York, 1, 1–88.

Bé, A.W.H., and W.H., Hamlin, 1967: Ecology of recent planktonic Foraminifera: Part 3. – Distribution in the North Atlantic during the summer of 1962: Micropaleontology, 13 (1), 87–106.

Behre, K-E., 1989: Biostratigraphy of the last glacial period in Europe. Quaternary Science Rew. 8, 25–44.

Berger, W.H., 1968: Planktonic foraminifera: selective solution and paleoclimatic interpretation. Deep-Sea Res., 15 ,(1), 31–43.

Berger, W.H., 1969: Ecologic patterns of living planktonic Foraminifera. Deep-Sea Res., 16, 1–24.

Berger, W.H., 1970: Biogenous deep-sea sediments: fractionation by deep-sea circulation. Geol. Soc. Amer. Bull. 81, 1385–1402.

Berger, W. H., 1973: Deep-sea carbonates: Pleistocene dissolution cycles. J. Foraminiferal Res., 3, 187–195.

Berger, W.H., 1976: Biogenous deep-sea sediments: Production, preservation and interpretation. In: Riley, J.P., and R. Chester (Eds): Chemical Oceanography, 5, Acad Press., London, 265–388.

Berger, W. H., 1977: Deep-sea carbonate and the deglaciation preservation spike in pteropods and foraminifera. Nature, 269, 301–304.

Berger, W.H., 1978: Sedimentation of deep-sea carbonate: maps and models of varations and fluctuations. J. Foram. Res. 8(4), 286–302.

Berger, W.H. 1992: Pacific carbonate cycles revisited: arguments for and against productivity control. *In:* Ishizaki, K., and T. Saito (Eds): Centenary of Japanese mircopaleotology. Terra Sci., Tokyo, 15–25.

Berger, W., 1994: On the origin of paleoceanography. AGU spring meeting May 23–27 1994, Baltimore, Abstract U21A-1.

Berger, W.H., and E.L. Winterer, 1974: Plate stratigraphy and the fluctuating carbonate line. *In:* K.J. Hsü and H. Jenkyns: Pelagic sediments on land and under the sea. Publ. Int. Ass. Sediment., 1, 11–48.

Berger, W.H., R.C. Finkel, J.S. Killingley and V. Marchig, 1983: Glacial-Holocene transition in deep-sea sediments: manganese-spike in the east-equatorial Pacific. Nature, 303, 231–233.

Berger, W.H., and E. Vincent,1986: Deep-sea carbonates: reading the carbon-isotope signal. Geol. Rundschau 22, 249–269.

Berger, W.H., and A. Spitzy, 1988: History of atmospheric CO_2 : constrains from the deep-sea record. Paleoceanography, 3(4), 401–411.

Berger, W. H. et al., 1993: Quaternary oxygen isotope record of pelagic foraminifers: Site 805, Ontong Java Plateau. Proc. Ocean Drilling Program, Sci. Res., 130, 363–379.

Berger, W. et al. 1994: Quaternary time scale for the Ontong Java Plateau: Milankovitch template for Ocean Drilling Program. Geology, 22, 463–467.

Björk, G., 1990: Studies of the ice conditions and the

vertical circulation in the upper Arctic Ocean. Depth. of Oceanography, Univ. of Göteborg, (thesis).

Bornmalm, L., 1988: Changes in the oxygen and carbon isotope composition during the last 150,000 years in cores from the northeastern Atlantric and the Alboran Sea. Dept. of Marine Geology, Univ. Göteborg

Brachert, T.C., and W-Ch., Dullo, 1990: Correlation of deep sea sediments and forereef carbonates in the Red Sea: an important clue for basin analysis. Marine Geology, 92, 255–267.

Bradley, W.H., 1938: Mediterranean sediments and Pleistocene sea levels. Science, 88, 376–379.

Broecker, W. S., M. Ewing and B.C. Heezen, 1960: Evidence for an abrupt change in climate close to 11,000 years ago. Amer. J. Sci. 258: 429–448.

Brotzen F., and Dinesen A., 1959: On the stratigraphy of some bottom sections from the central Pacific. Rept. Swedish Deep-Sea Exped. 1947–1948, X (IV), 43–55.

Burckle, L. H. and R. B., McLaughlin, 1977: Size changes in the diatom *Coscinodiscus nodulifer* A. Schmidt in the equatorial Pacific. Micropaleontology, 23, (2), 216–222.

Burckle, L. H., N. J. Shackleton, and S. L. Bromble, 1981: Late Quaternary stratigraphy for the equatorial Pacific based upon the diatom *Coscinodiscus nodulifer*. Micropaleontology, 27, (4), 352–355.

Butzer, K.W., 1964: Environment and archeology. An introduction to Pleistocene geography. Aldine Publ. Co., Chicago, 524 pp.

Cita, M.B., et al., 1977: Paleoclimatic record of a long deep-sea cores from the eastern Mediterranean. Quat. Res., 8, 205–235.

Cita, M. B., and Ryan, W. B. F., 1973: The Pliocene Record in deep sea Mediterranean sediments. V. Time-scale and general synthesis. Init. Repts, Deep Sea Driiling Program, 13, 2 U.S. Government Printing Office, Washington, D.C., 1405–1416.

Crane, R.G., and R.G. Barry, 1984: The influence of clouds on climate with a focus on high latitude interactions. J. Climatol. 4 (1) 71–93.

Dansgaard, W., S.J. Johnsen, H.B. Clausen, and C.C. Langway, Jr., 1971: Climatic record revealed by Camp Century ice core. *In:* K.K. Turekian (Ed): Late Cenozoic glacial ages. Yale Univ. Press, Hartford, 37–56.

Dansgaard, W., et al., 1993: Evidence for general instability of past climate from a 250-kyr ice-core record. Nature, 364, 218–220.

Defant, A., 1961: Physical oceanography. Pergamon Press, London.

Degens, E.T., How King Wong, and M.G. Wiesner, 1986: The Black Sea region: sedimentary facies, tectonics and oil potential. Mitt. Geol.-Paläont. Inst. Univ. Hamburg, Heft 60, 127–149.

Degens, E. T. and D. A. Ross (ed): 1969: Hot brines and recent heavy metal deposits in the Red Sea. A geochemical and geophysical account. Springer-Verlag, Berlin, 600 pp.

Delmas, R.J., Ascenio, J.M. and Legrand, M. 1980: Polar ice evidence that the atmospheric CO_2 20,000 years B.P. was 50% of the present. Nature, 284, 155–157.

Deuser, W.G., and E.T., Degens, 1969: O^{18}/O^{16} and C^{13}/C^{12} ratios of fossils from the hot-brine deep area of the central Red Sea: *In:* E.T. Degens and D.A. Ross (Eds.): Hot brines and recent heavy metal deposits in the Red Sea. Springer, Berlin, 336–347.

Duplaix, S., 1958: Etude Mineralogique des niveaux sableux des carottes prélevées sur le fond de la Méditerranée. Rept. Swedish Deep-sea Exped. 1947–1948, VIII, 2(1), 139–166.

Ehlers, J., 1983: The glacial history of the north-west Germany. *In:* J. Ehlers (Ed.): Glacial deposits in North-West Europe. Balkema, Rotterdam, 229–238.

Emiliani, C., 1954: Depth habitats of some species of pelagic foraminifera as indicated by oxygen isotopic ratios. Amer. J. Sci., 252, 149–158.

Emiliani, C., 1955 a: Pleistocene temperatures. J. Geol. 63, 538–578.

Emiliani, C., 1955 b: Pleistocene temperature variations in the Mediterranean. Quaternaria, 2, 87–98.

Emiliani, C. 1956: Oligocene and Miocene temperatures of the equatorial and subtropical Atlantic Oceans. J: Geol. 64 (3), 281–288.

Emiliani, C., 1958: Paleotemperature analysis of Core 280 and Pleistocene correlations. J. Geol. 66, 264–275.

Emiliani, C., 1978: The cause of the ice ages. Earth. Planet. Sci. Lett., 37, 349–352.

Epstein, S. and T. Mayeda, 1953: Variation of O^{18} content of waters from natural sources. Geochim. et Cosmochim. Acta, 4, 213–224.

Ericson, D.B. and G. Wollin, 1956: Correlation of six cores from the equatorial Atlantic and the Caribbean. Deep-Sea Res. 3, 104–125.

Ericson, D.B., M. Ewing, G. Wollin, 1964: Sediment cores from the Arctic and Subarctic Seas. Science, 144 (3623), 105–116.

Eriksson, K.G., 1965: The sediment core no. 210 from the western Mediterranean Sea. Rept. Swedish Deep-Sea Exped., 1947–1948, VIII, 5, 393–594, 26 plates.

Farrell, J.W., and W.L. Prell, 1991: Pacific $CaCO_3$ preservation and $\partial^{18}0$ since 4 Ma: paleoceanic and paleoclimatic implications. Paleoceanography, 6 (4), 485–498.

Flint, R.F., 1949: Glacial geology and the Pleistocene epoch. J. Wiley & Sons, New York, 589 pp.

Flint, R.F., 1971: Glacial and Quaternary Geology. J. Wiley, New York, 892 pp.

Flohn, H, 1969: Ein geophysicalisches Eiszeit-Modell. Eiszeitalter u. Gegenwart, 20, 204–231.

Forsström, L, and M. Eronen, 1985: Flandrian and Eemian shore levels in Finland and adjacent areas – a discussion. Eiszeitalter u. Gegenwart, 35, 135–145.

Fredriksson, K., 1959: A sediment core from the Mindanao Trough. Rept. Swedish Deep-Sea Exped. 1947–1948, VI (4), 61 p. and 14 plates.

Goel, P. S., D. P., Kharkar, D. Lal, N., Narseppaya, B., Peters, and V. Yatirajam, 1957: The beryllium -10 concentration in deep-sea sediments. Deep-Sea Research, 4, 202–210.

Goldberg, E.D., and M. Koide, 1961: Geochronological studies of deep-sea sediments by the ionium-thorium method. Geochim. Cosmochim. Acta. 26, 417–450.

Gordon, A., 1986: Is there a global scale oceanic circulation? Eos, 67,109–110.

Gray, C.G., et al., 1992: Geochronology and subsurface stratigraphy of Pukapuka and Rakahanga atolls, Cook Islands: Late Quaternary reef growth and sea level history. Palaeogeogr., Palaeoclimatol., Palaeoecol., 91, 377–394.

Grosswald, M.G., 1980: Late Weichselian ice sheet of the Northern Eurasia. Quat. Res., 13, 1–32.

Haupt, B.J., C. Schäfer-Nethn, and K. Stattegger, 1994: Modeling sediment drifts: a coupled oceanic circulation – sedimentation model of the northern North Atlantic. Paleoceanography, 9, (6), 897–916.

Hays, J. D., J. Imbrie, and N. J. Shackleton, 1976: Variations in the Earth´s orbit: pacemaker of the ice ages. Science, 194, 1121–1132.

Hemleben, C., 1986: Environmental limitations and physiological requirements of planktonic foraminifers. *In:* Second Int. Conf. Paleoceanography, Woods Hole, Mass.

Hodell, D. A., D.F. Williams, and J. P. Kennett, 1985: Late Pliocene reorganisation of deep vertical water-mass structure in the western South Atlantic: Faunal and isotopic evidence. Geol. Soc. Am. Bull. 96, 495–503.

Hsu, K. J., et al. 1977: History of the Mediterranean salinity crisis. Nature, 267, 399–403.

Imbrie J. and N.G., Kipp, 1971: A new micropaleontological method for quantitative paleoclimatology: application to a late Pleistocene Caribbean core. *In:* K.K., Turekian (Ed): The late Cenozoic ages, Yale Univ. Press, Hartford, 71–182.

Jansen, J. H. F., A. Kuijpers, and S. Troelstra, 1986: A Mid-Brunhes climatic event: Longterm changes in global atmosphere and ocean circulation. Science, 232, 619–622.

Johnsen, S.J. et al., 1992: Irregular glacial interstadials recorded in a new Greenland ice core. Nature , 359, 311–313.

Joly, J., 1908: On the radium content of deep-sea sediments. Phil. Mag., 6, 190.

Jouzel, J. et al., 1993: Extending the Vostok ice-core record of palaeoclimate to the penultimate glacial period. Nature, 364, 407–412.

Kennett, J. 1982: Marine Geology. Prentice. Hall, Englewood Cliffs, 813 pp.

Kidd, R., M.B. Cita, and W.B.F. Ryan, 1978: Stratigraphy of the eastern Mediterranean sapropel sequences recovered during DSDP leg 42A and their paleoenvironmental significance. *In:* K. Hsü, L. Montadert et al: Initial Reports of the Deep Sea Drilling Project, 42, 1, Washington (U.S. Government Printing Office), 421–443.

Kolbe, R.W. 1954: Diatoms from Equatrial Pacific cores. Rept. Swedish Deep-Sea Exped. 1947–1947, VI, fasc. 1. 49 pp., 4 plates.

Kolbe, R.W., 1955; Diatoms from Equatorial Atlantic cores. Ibid. VII, fasc. 3, 50 pp, 4 plates.

Kolbe, R. W. 1957: Diatoms from Equatorial Indian Ocean cores. Ibid. IX, fasc. 1, 38 pp, 2 plates.

Kuenen, Ph. H., 1950: Marine Geology, J. Wiley & Sons, New York, 568 pp.

Kuijpers, A., 1989: Southern ocean circulation and global climate in the middle Pleistocene (early Brunhes). Palaeogeogr., Palaeoclimatol., Palaeoecol., 76, 67–83.

Kullenberg, B., 1944: Kolvlodet. Geol. För. Förh. Stockholm, 66, 501.

Kullenberg, B., 1947: The piston core sampler. Svenska hydr.-biol. komm. skr. Tredje ser.: Hydr. Göteborg.1 (1).

Kullenberg, B., 1952: On the salinity of the water contained in marine sediments. Medd. Oceanografiska Inst. Göteborg 21, 38 pp.

Kullenberg, B., 1953: Absolut chronology of deep-sea sediments and the deposition of clay on the ocean floor. Tellus, 5 (3), 302–305.

Kullenberg, B., 1955: Deep-Sea coring. Rept. Swedish Deep Sea Exped. 1947–1948, IV (1), 35–96.

Labeyrie, L. D., et al., 1986: Melting history of Antarctica during the past 60 000 years. Nature, 322, 701–706.

Lange, C. B. and W. H. Berger, 1993: Diatom productivity and preservation in the western equatorial Pacific: the Quaternary record.. Proc. Ocean Drilling Program, Scientific Res. 130, 509–523.

Lisitzin, A.P., 1985: The silica cycle during the last ice age. Paleogeogr., Palaeoclim., Palaeoecol., 50 (2/3), 241–270.

Locke, S., and R.C. Thunell, 1988: Paleoceanographic record of the last glacial/interglacial cycle in the Red Sea and Gulf of Aden. Palaeogeogr., Palaeoclimatol., Palaeoecol., 64, 163–187.

Lorius, C., et al., 1979: A 30 000-yr isotope climatic record from Antarctic ice. Nature, 280, 644–648.

Lorius, C., et al., 1988: Antarctic ice core: CO_2 and climate change over the last climatic cycle. EOS, June 28, 681–684.

Lunkad, S.K., 1981: Late Quaternary anoxic sediments of the eastern Mediterranean Sea: mineralogical, microfaunal and isotopic study. Doctoral thesis, Univ. Pierre et Marie Curie, Paris, no. 81–50, 261 pp.

Markussen, B., R. Zahn and J. Thiede, 1985: Late Quaternary sedimentation in the eastern Arctic Basin: stratigraphy and depositional environment. Palaeogeograpgy, Palaeoclimatology, Palaeoecology, 50 (2/3), 271–284.

McCoy, F. W., 1974: Late Quaternary sedimentation in the eastern Mediterranean Sea. Ph. D. Diss. Harward University.

McCrea, J.M. 1950: On the isotope chemistry of carbonates and a paleotemperature scale. J. Chem. Physics, 18 849–857.

McIntyre, A., N. G. Kipp, A. W. H. Bé, T. Crowley, T. B. Kellogg, J. V. Gardner, W. Prell and W. Ruddiman, 1976: Glacial North Atlantic 18,000 years ago: A CLIMAP reconstruction. Geol. Soc. Amer. Memoir, 145, 43–76.

Mellis, O., 1954: Volcanic ash-horizons in deep-sea sediments. Deep-Sea Res., 2, 89–92.

Mikhailescu, C., and S. Medianic, 1994: Quaternary glacial-interglacial oscillations in the Black Sea region. Am Quatern. Assoc., Abstract, 19–22 June 1994, Univ. Minnesota, 134–136.

Mikkelsen, N., 1977: On the origin of Ethmodiscus ooze. Marine Micropaleontology, 2, 35–46.

Miller, K.G., Fairbanks, R.G., and Thomas, E. 1987: Benthic foraminiferal carbon isotopic records and the development of abyssal circulation in the eastern North Atlantic. *In:* W.F. Ruddiman, R.B. Kidd, E. Thomas et al., Initial Reports of the Deep Sea Drilling Project, 94, U.S. Goverment Printing Office, Washington, D.C., 981–996.

Mix, A.C., and W.F. Ruddiman, 1984: Oxygen-isotope analyses and Pleistocene ice volumes. Quaternary Res. 21, 1–20.

Mosby, H., 1962: Water, salt and heat balance of the North Polar Sea and the Norwegian Sea. Geophysica Norwegica 24, (11), 289–313.

Murray, J. and A.F. Renard, 1894: Rep. Scient. Res. H.M.S. "Challenger" 1873–76, IV.

Müller, H; 1974: Pollenanalytische Untersuchungen und Jahresschichtenzählungen an der holsteinzeitliche Kiesel-gur von Munster. Breloch. Geologisches Jahrbuch A 21, 107–140.

Naidu, P.D., B.A. Malmgren and L. Bornmalm, 1993: Quaternary history of calcium carbonate fluctuations in the western equatorial Indian Ocean (Somali Basin). Palaeogeography, Palaeoclimatology, Palaeoecology, 103, 21–30.

Neftel, A., Oeschger, H., Schwander, J., Stauffer, B., and Zumbrunn, R., 1982: Ice core sample measurements give atmospheric CO_2 content during the past 40,000 yrs. Nature, 295, 220–223.

Nier, A.O. 1947: A mass spectrometer for isotope and gas analysis. Rev. Sci. Instruments 18, 398–411.

Nilsson E., 1938: Pluvial lakes in East Africa. GFF, Stockholm, 60, 423–433.

Ninkovich, D. and B.C. Heezen, 1965: Santorin Tephra. *In:* W.F. Whittard and R. Bradshaw

(Eds): Submarine geology and geophysics, London, 413–453.

Norin, E., 1958: The sediments of the central Tyrrhenian Sea. Rept. Swedish Deep-Sea Exped. 1947–1948, VIII, 1, 1–136, 19 plates.

Olausson, E., 1957: Das Moor Roshultsmyren. Eine geologische, botanische and hydrologische studie in einem Südwestschwedischen Moor mit exzentrich gewölbten Mooselementen. Lunds Univ. årsskrift, N.F., avd. 2, bd 53, (12), 72 pp.

Olausson, E., 1960 a: Description of sediment cors from the Central and Western Pacific with the Adjacent Indonesian Region. Rept. Swedish Deep-Sea Exp. 1947–1948, VI (V) 161–214. 42 plates.

Olausson, E., 1960 b: Description of sediment cores from the Indian Ocean, Ibid., IX (II), 51–88, 18 plates

Olausson, E., 1960 c: Description of sediment cores from the Mediterranean. Ibid. VIII (III), 285–334, 24 plates.

Olausson, E., 1960 d: Description of sediment cores from the North Atlantic. Ibid. VII (V), 227–286, 42 plates.

Olausson, E., 1961 a: Studies of deep-sea cores. Ibid. VIII (IV), 335–391,2 plates.

Olausson E., 1961 b: Remarks on some Cenozic sequences from the central Pacific, with a discussion of the rôle of coccolithophorids and foraminifera in carbonate deposition. Medd. Oceanogriska institutionen, Göteborg 29. 35 pp.

Olausson, E., 1961 c: Remarks on two Tertiary sequences of two cores from the Pacific. Bull. Geol. Inst. Univ. Uppsala, XL, 299–303.

Olausson, E., 1965: Evidence of climatic changes in North Atlantic cores, with a remarks on isotopic paleotemperature analysis. Progress in Oceanography 3, Pergamon, New York, 221–252.

Olausson, E. , 1966: Sediments of Atlantic. Sediments of Pacific. Sediments of Mediterranean. *In:* Rh.W. Fairbridge (Ed.): The Encyclopedia of Oceanography. Reinhold, New York, 75–80, 493–497, 665–669.

Olausson, E., 1967: Climatological, geoeconomical and paleooceanographical aspects on carbonate deposition. Progress in Oceanography, 4, Pergamon, New York, 245–265.

Olausson, E., 1969: On the Würm-Flandrian boundary in deep-sea cores. Geol. Mijnbouw, 48, 349–361.

Olausson, E., 1970: E.T. Degens and D. A. Ross (Eds.): Hot brines and recent heavy metal deposits in the Red Sea. Springer, Berlin, 1969, 600 pp. – a review. GFF, 92 (2), Stockholm, p. 247.

Olausson, E., 1970 b: Water sediment exchange and recycling of pollutants through biogeochemical processes. *In:* FAO technical conference on marine pollution and its effect on living resources and fishing. FIR:MP/R-31, 23 Nov. 1970, Rome, 1–18.

Olausson, E., 1971 a: Quaternary correlations and the geochemistry of oozes. *In:* Funnell, B.M. and W. Riedel (Eds): The Micropaleontology of oceans. Cambridge Univ. Press, 375–398.

Olausson, E., 1971 b: Tephrochronology and the Late Pleistocene of the Aegean Sea. Opera Botanica 30, 29–39.

Olausson, E., 1972 a: Norwegian Sea in an ice age model. Ambio spec. rep. 2, 13–17.

Olausson, E., 1972 b: Oceanographic aspects of the Pleistocene of the Arctic Ocean. Inter-Nord 12, 151–170.

Olausson, E., 1975: Man-made effect on sediments from Kattegat and Skagerrak. Geol. För. Förh., Stockholm, 97, 3–12.

Olausson, E., 1981 a: On the glacial Norwegian – Greenland Seas and the Arctic Ocean. Quaternary Studies in Poland, 3, 79–89.

Olausson, E., 1981 b: On the isotopic composition of Late Cenozoic sea water. Geograf. Annaler, 63 A, 3–4.

Olausson, E., 1982: General comments on the Pleistocene/Holocene boundary. *In:* Olausson, E., (Ed.): The Pleistocene/Holocene boundary in south-western Sweden. Sv. Geol. Unders., C 794, Uppsala, 10–15.

Olausson, E., 1982 b: Pleistocene stratigraphy and chronology. GFF, 104, 255–256.

Olausson, E., 1983: On the Late Pleistocene exchange of water across the the Icelandic transverse ridge. *In:* M.H.P. Bott, S. Saxov, M. Talwani and J. Thiede (Eds.): Structure and development of the Greenland-Scotland Ridge. Nato Conf. Ser. IV: Marine Sciences, 8, Plenum Press, New York, 591–599.

Olausson E., 1985: The glacial oceans. Palaeogeography, Palaeoclimatology, Palaeoecology, 50 (2/3), 291–301.

Olausson, E., 1988: On the Cenozoic oceans: evidence of the calcium carbonate content, $\partial^{13}C$ and $\partial^{18}0$. Ibid. 67 (1/2), 103–118.

Olausson, E., 1991 a: Carbon and oxygen isotope composition of foraminifera in two cores from the Bannock Basin area, eastern Mediterranean. Marine Geology, 100 (1/4), 45–51.

Olausson, E., 1991 b: A post-Cromerian rise in sea-level. Int. conf. on the rôle of the polar regions in global change. Proceed. Conf. June 11–15, 1990, Univ. of Alaska, Fairbridge, 496–498.

Olausson, E., 1995: The last three northern deglaciations in Eurasia, particularly the Elsterian/Holsteinian transition. Eyjar i Eldhafi. Gott mál Hf, Reykjavik, 199–207.

Olausson, E., and S. Uusitalo, 1963: On the seismic vibrations on sediments. Comptes Rendes de la Soc. géologique de Finlande XXXV, 101–113

Olausson, E., and U.C. Jonasson, 1969: The Arctic Ocean during the Würm and early Flandrian. Geol. Fören. Stockholm Förh. 91, 185–200.

Olausson, E., and I.U. Olsson, 1969: Varve stratigraphy in a core from the Gulf of Aden. Paleogeogr., Palaeoclimatol., Palaeoecol., 6, 87–103.

Olausson, E., U. Z. Bilal Ul Haq, G. B. Karlsson and I. U. Olsson 1971: Evidence in Indian Ocean cores of Late Pleistocene changes in oceanic and atmospheric circulation. Geol. För. Förh. Stockholm, 93 (1), 51–84.

Olausson, E., and B. Svenonius, 1975: Past changes in the geomagnetic field caused by glaciations and deglaciations. Boreas, 4, 55–62.

Olausson, E., and I. Cato (ed.),1980: Chemistry and biogeochemistry of estuaries. J. Wiley & Sons Ltd, Bath, 452 pp.

Olsson J., 1925: Väg- och vattenbyggnadskonst. Tekn. tidskr., Stockholm 55, 13.

Parisi, E., 1987: Carbon and oxygen isotope composition of *Globigerinoides ruber* in two deep-sea cores from the Levantine basin (Eastern Mediterranean). Marine Geology, 75, 201–219.

Parker, F.L., 1958: Eastern Mediterranean Foraminifera: Rept. Swedish Deep-Sea Exped. 1947–1948, VIII 2(4),219–283, 16 diagrams and 6 plates.

Pettersson, H., 1937: The proportion of thorium to uranium in rocks and in the sea. Anz. Akad. Wiss. Wien, Math. Naturwisse. Klass127–128.

Pettersson, H., 1950: Med Albatross över havsdjupen. Bonniers, Stockholm, 239 pp.

Pettersson, H., 1955: Shooting stars. *In:* Sjöfartsmuseets in Gothenburgd Festschrift to Tor E. J:son Broström, Göteborg.117–132.

Pettersson, H., 1957: The voyage. Rept. Swedish Deep-Sea Exped. 1947–1948, I (1), 124 pp, 2 plates.

Pettersson, H. 1958: Rate of accretion of cosmic dust on the Earth. Nature, 181,330.

Pettersson, H., and K. Fredriksson, 1958: Magnetic spherules in deep-sea deposits. Pacific Science, 12, 71–81.

Pettersson, H., and B. Kullenberg 1941: Vacuumlodet. Medd. Oceanografiska Inst. Göteborg, 5.

Petersson, H., et. al., 1948: Three sediment cores from Tyrrhenian Sea. Ibid. 15.

Phleger, F. B. Jr., 1948: Foraminifera of a submarine core from the Caribbean. Ibid. 16.

Phleger, F. B., F. L. Parker and J.F. Peirson 1953: North Atlantic foraminifera. Rept. Swed. Deep-Sea Exped., 1947–1948, III (1), 121 pp.

Raymo, M. E., W. F. Ruddiman, N .I. Shackleton and D. W. Oppo, 1990: Evolution of Atlantic – Pacific $\partial^{13}C$ gradients over the last 2.5 m.y. Earth and Planetary Sci. Lett., 97, 353–368.

Revelle, R. R., 1944: Marine bottom samples collected in the Pacific Ocean by the Carnegie on its seventh cruise. Carnegie Inst. Wash. 556 (1), 180 pp.

Riedel, W. R., 1957: Radiolaria. A. Preliminary stratigraphy. Rept. Swedish Deep-Sea Exped. 1947–1948, VI (III), 61–96, 4 plates

Robin. G. de Q, 1988: The Antarctic ice sheet, its history and response to sea level and climatic changes over the past 100 million years. Paleogeogr., Palaoeoclimatol., Palaeoecol, 67, 31–50.

Rosholt, J.N., C. Emiliani, J. Geiss, F.F. Koczy, and P.J. Wangersky, 1961: Absolute dating of deep-sea cores by the Pa^{231}/Th^{230} method. J. Geol. 69 (2),161–185.

Rossignol-Strick, M. W., et al., 1982: After the deluge: Mediterranean stagnation and sapropel formation. Nature, 295, 105–110.

Ruddiman, W.F., A. McIntyre, 1981: the North Atlantic Ocean during the Last deglaciation. Palaeogeogr., Palaeoclim., Palaeoecol., 35, 145–214.

Ruddiman, W F., et. al., 1987: Paleoenvironmental results from North Atlantic Sites 607 and 609. *In:* W.F. Ruddiman et al.: Initial Reports of the Deep Sea Drilling Project, 94. U.S. Government Printing Office, Washington, D.C., 855–878.

Ruddiman, W. F., and J.E. Kutzbach, 1989: Forcing of

late Cenozoic northern hemisphere climate by plateau uplift the southern Asia and the American west. J. Geophys. Res. 94, D15, 18409–18427.

Salinger, M. J., 1981: Paleoclimates north and south. Nature 291, 106–107.

Sanyal, A., N.G. Hemming, G.N. Hanson, and W.S. Broecker, 1995: Evidence for a higher pH in the glacial ocean from boron isotopes in foraminifera. Nature, 373, 234–236.

Schott, W., 1935: Die Foraminiferen in dem aequatorialen Teil des Atlantischen Ozeans. Wiss. Ergebn. deutsch. Atl. Esp. Forsch. u. Verm. "Meteor", III (3), 43–134.

Schott, W., 1952: On the sequeence of deposits in the equatorial Atlantic Ocean. Göteborgs K. Vetensk. Vit. Samh. Handl. Sjätte följden, B, 6 (2), 15 pp.

Schott, W., 1966: Foraminiferenfauna und Stratigraphie der Tiefsee. sedimente im Nordatlantischen Ozean. Rept. Swedish Deep-Sea Exped. 1947–1948, VII , 8, 357–468.

Sclater , J. G et al., 1985: The depth pf the ocean through the Neogene. *In:* Kennett, J. P. (Ed.): The Miocene Ocean. Geol. Soc. Amer. Mem. 163, 1–19.

Schrader, H. J., 1978: Quaternary through Neogene history of the Black Sea, deduced from the paleoecology of diatoms, silicoflagellates, ebridians and chrysomonads. *In:* D. A. Ross, Y.P. Neprochnov, et al.: Initial Reports of the Deep Sea Drilling Project 42 (2), Washington (U.S. Government Printing Office), 789–867.

Sclater, J.G., L. Meinke, A. Bennett, and Murphy, C., 1985: The depth of the ocean through the Neogene. Geol. Soc. Amer. Memoir, 163, 1–19.

Shackleton N., and N. D. Opdyke, 1976: Oxygen-isotope and paleomagnetic stratigraphy of equatorial Pacific Core V28-239 late Pliocene to latest Pleistocene. *In:* R. M., Cline and Hays, J. D. (Eds.): Investigation on late Quaternary paleoceanography and paleoclimatology. Geol. Soc. Am. Mem. 145, 449–464.

Shackleton, N.I. and M.A. Hall, 1984: Oxygen and carbon isotope stratigraphy of deep sea drilling project Hole 552A: Plio-Pleistocene glacial history. *In:* D.G. Roberts, D. Schnither et al., Initial Reports of the Deep Sea Drilling Project, 81, U.S. Goverment Printing Office, Washington, D.C., 599–610.

Slingo, A., 1990: Sensitivity of the Earth´s radiation budget to changes in low clouds. Nature, 343, 49–51.

Stigebrandt, A., 1985: In the hydrographic and ice conditions in the northern North Atlantic during different phases of a glacial cycle. Palaeogeogr., Palaeoclimatol., Palaeoecol. 50 (2/3), 303–322.

Stradner, H., 1973: Catalogue of calcareous nannoplankton from sediments of Neogene age in the eastern North Atlantic and the Mediterranean Sea. Init. Rept. Deep-Sea Drilling Project, 13, U.S. Government Printing Office, Washington, D.C., 1137–1199.

Sverdrup, H. U., M. W. Johnson and R. Fleming 1942: The Oceans. Prentice-Hall, New York, 1087 pp.

Thiede, J. 1978: A glacial Mediterranean. Nature 276, 680–683.

Thierstein, H.R., et al. 1977: Global synchroneity of late Quaternary coccolith datums: validation by oxygen isotopes. Geology, 5, 400–404.

Thunell, R.C., and D.F. Williams, 1983 a: Paleotemperature and paleosalinity in the eastern Mediterranean during the Quaternary. Palaeogeogr., Palaeoclimatol., Palaeoecol., 44, 23–39.

Thunell, R.C., D.F. Williams, and M.B. Cita, 1983 b: Glacial anoxia in eastern Mediterranean, J. Foraminifer. Res. 13, 283–290.

Thunell, R. C., D.F. Williams and M. Howell, 1987: Atlantic-Mediterranean water exchange during the late Neogene. Paleoceanography, 2 (6), 661–678.

Todd. R., 1958: Foraminifera from western Mediterranean cores. Rept. Swedish Deep-Sea Exped., 1947–1948, VIII, 2 (3),169–215, 19 tables.

Urey, H.C., 1947: The thermodynamic properties of isotopic substances. J. Chem. Soc. 562–581.

Utech, K., 1963: Die Konzentration kosmischer Kügelchen in Sedimenten als Massstab für deren Bildungsgeschwindigkeit. Neues Jahrb. Geol. Palaeontol. Abhandl. 116, 325–333.

Walin, G., 1985: The thermohaline circulation and the control of ice ages. Paleogeogr., Palaeoclim., Palaeoecol., 50 (2/3), 323–332.

van Andel, Tj. H., G.R. Heath and T.C. Moore, 1975: Cenozoic tectonics, sedimentation, and paleoceanography of the central equatorial Pacific. Geol. Soc. Am. Memoir 143, 134 pp.

Van der Hammen, T., T.A. Wijmstra, and W.H. Zagwijn, 1971: The floral record of the late Cenozoic of Europe. *In:* K. K. Turekian (Ed.);

Late Cenozoic glacial ages. Yale Univ. Press, Hartford, 391–424.

Weyl, P.K., 1968: The rôle of the oceans in climatic change: a theory of the ice ages. Meteorol. Monogr., 8 (30), 37–62.

Whitman J. M., and W. H. Berger, 1992: Pliocene-Pleistocene oxygen isotope record Site 586, Ontong Java Plateau. Mar. Micropaleontol., 18, 171–198.

Vincent, E., J. S. Killingley, and W.H. Berger, 1985: Miocene oxygen and carbon isotope stratigraphy of the tropical Indian Ocean. *In:* J. P. Kennett (ed.): The Miocene Ocean. Geol. Soc. Am. Mem., 163, 103–130.

Wiseman J. D. H., 1965: The changing rate of calcium carbonate sedimentation on the equatorial Atlantic floor and its relation to continental late Quaternary stratigraphy. Rept. Swedish Deep-Sea Exped. 1947–1948. VII, fasc. 6 , 287–354, 4 plates.

Woldstedt, P., 1969: Quartär. Handbuch der Stratigraphischen Geologie Bd II. Ferdinand Enke Verlag, Stuttgart, 263 pp.

Woodruff, F., S. M. Savin, and R.G. Douglas, 1981: Miocene stable isotope records: a detailed deep Pacific Ocean study and its paleoclimatic implications. Science, 212, 665–668.

Wu, G., M.K. Yasuda and W. Berger, 1991: Late Pleistocene carbonate stratigraphy on Ontong-Java Plateau in the western equatorial Pacific. Marine Geology, 99: 135–150.

Zagwijn, W.H., 1974: The Pliocene- Pleistocene boundary in western and southern Europe. Boreas, 3, 75–97.

Zubakov, V.A. and I.I. Borzenkova, 1990: Global paleoclimate and the later Cenoic. Developments in Palaeontology and Stratigraphy, 12, Elsevier, Amsterdam, 456 pp.

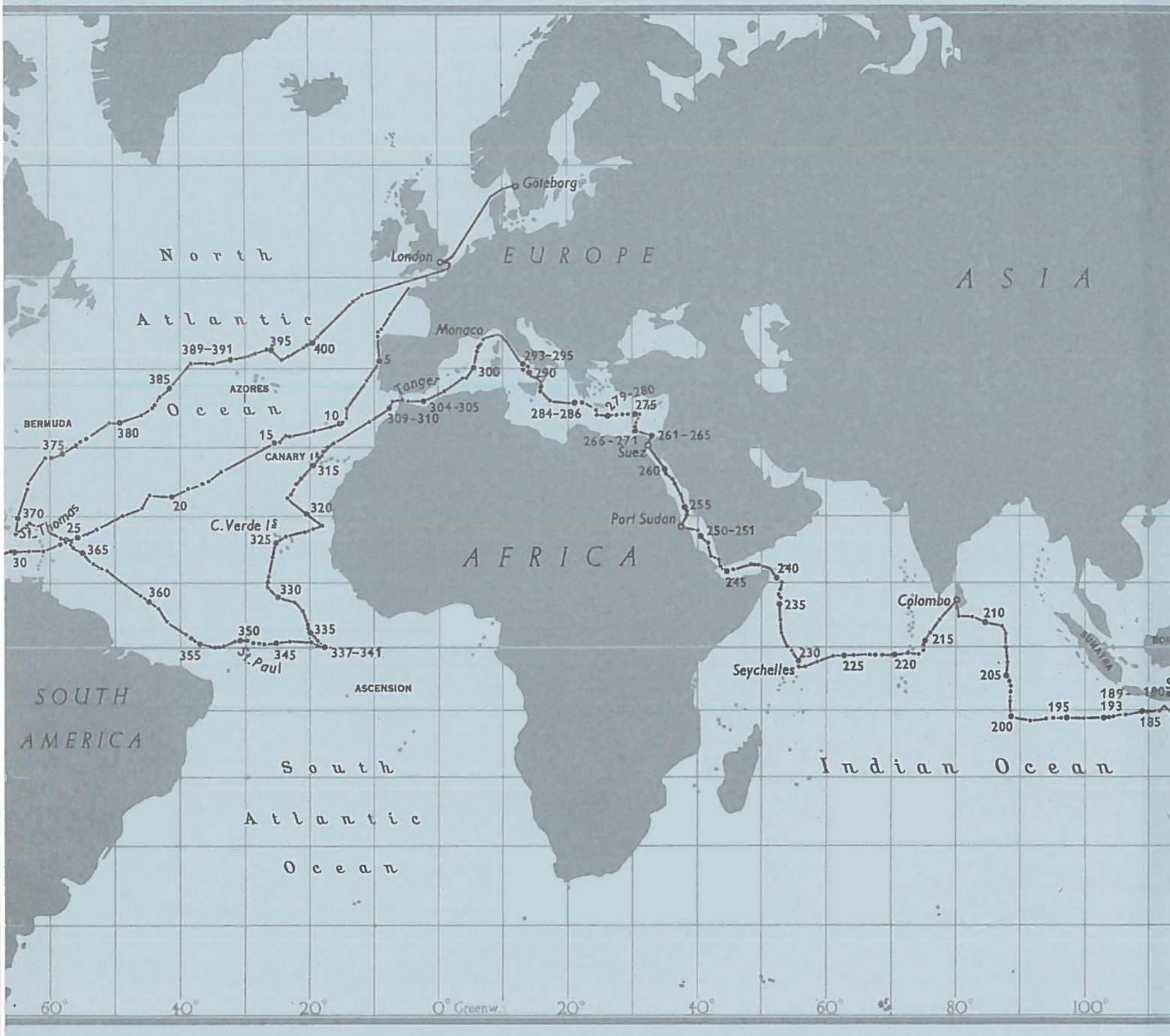
Four famous scientists mentioned above

Dr Cesare Emiliani (1922–1995).

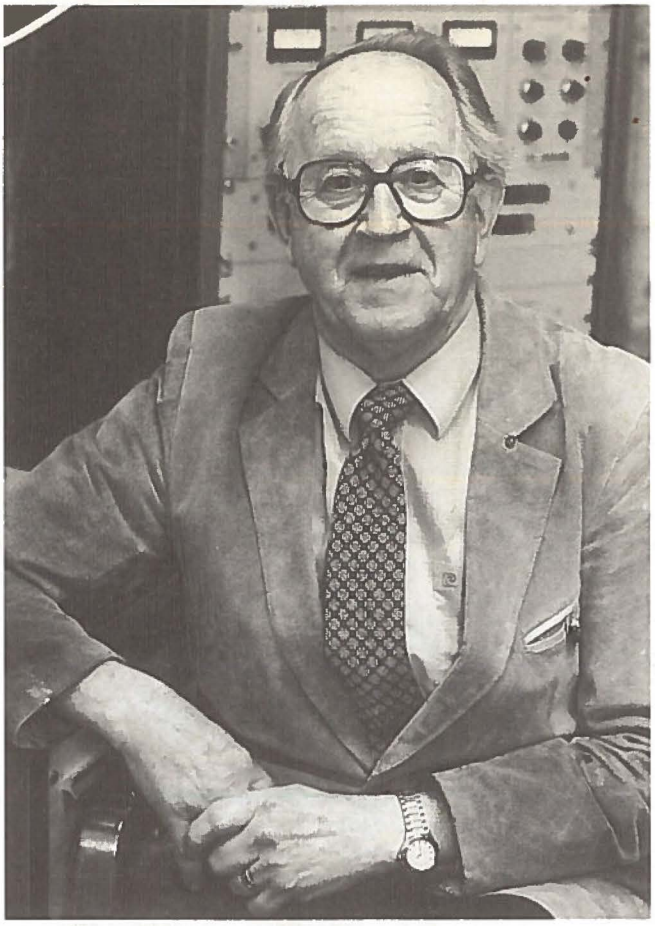
Dr Fred Phleger (to the left) and Dr Hans Pettersson at a congress.

Dr Gustaf Arrhenius.

THE TRACK OF MS



ALBATROSS 1947 - 1948



Professor Eric Olausson
(Foto: Kamerareportage, by permission)

On July 4 1997 it is fifty years since the "Albatross" left Göteborg on its fifteen months long deep sea expedition around the world. During this cruise the scientists on board collected ca 400 long sediment cores with a maximum length of 20 m. The studies of these cores bridged the gap between oceanography and geology. A new science was born: paleoceanography. Instead of studying the ocean's processes on a time scale of less than a thousand years, the perspective was extended approx two million years. The complicated interplays between the oceans and their adjacent seas were discovered and found to mirror and influence the climatic changes. By combining these records from the Albatross cores with those obtained from studies of continental deposits, particularly from present and past glaciated areas, a new picture of the Quaternary epoch was established.

It is the aim of this book to sum up the results of the study of the Albatross cores on the occassion of this anniversary celebration and to remind scientists and other people interested in marine science about important Swedish contributions to the study of past oceans.